Volume 11, Number 1
January–February 2003

ISSN: 0869-5938

STRATIGRAPHY AND GEOLOGICAL CORRELATION

English Translation of *Stratigrafiya. Geologicheskaya Korrelyatsiya*

Editor-in-Chief
Boris S. Sokolov

http://www.maik.ru

Journal of the Fundamental and Applied Aspects of Stratigraphy and the Correlation of Geologic Events and Processes in Time and Space

МАИК "НАУКА/INTERPERIODICA" PUBLISHING

Stratigraphy and Geological Correlation

(Stratigrafiya. Geologicheskaya Korrelyatsiya)

ISSN: 0869-5938

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PUBLISHER
INTERNATIONAL ACADEMIC PUBLISHING
COMPANY "NAUKA/INTERPERIODICA"
PROFSOYUZNAYA UL. 90, MOSCOW, 117997 RUSSIA
PHONE: 7 (095) 336-1600; FAX: 7 (095) 336-0666
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SUBSCRIPTION PRICES (2003, Volume 11, 6 issues)

US $1270.00 (North America), $1455.00 (elsewhere). Subscriptions may be entered at any time. Orders may be placed at agencies or directly. Cancellations not allowed except for duplicate payment. Prepayment required. Advance payment required in US$/US bank. MasterCard/Visa accepted. Make checks payable to *Stratigraphy and Geological Correlation* or Interperiodica.

Stratigraphy and Geological Correlation (ISSN: 0869-5938) is published bimonthly for US $1270.00 (North America), outside North America for $1455.00 per year by MAIK "Nauka/Interperiodica", 5724 Hwy. 280 East, Birmingham, Alabama 35242-5724 USA. Periodical postage paid at Birmingham, Alabama, and additional mailing offices. POSTMASTER: Send address changes to *Stratigraphy and Geological Correlation*, P.O. Box 1831, Birmingham, Alabama 35201 USA.

The journal was founded in 1993.

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Printed in the USA by Pleiades Publishing, Inc.

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Simultaneous English language translation of the journal is available from MAIK "Nauka/Interperiodica" (Russia).
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Stratotype of the Famennian Volgograd Horizon in the Russian Platform

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Received June 20, 2000

Abstract—Stratotype sections of the lower Famennian Volgograd Horizon, which are located in the Umet–Linevo paleodepression, the northeastern Volgograd area, are described. Lithological–geophysical characteristics of borehole sequences, correlation and paleogeological profile of the Frasnian–Famennian boundary beds of the Volgograd region are presented. Stratigraphic ranges of characteristic spore species of the Volgograd Horizon are defined. Palynological data elucidate correlation of the Frasnian–Famennian boundary beds in the Russian platform and the Urals. Distribution areas of the Volgograd Horizon deposits in the Russian platform are outlined.

Key words: Famennian Stage, Volgograd Horizon, Umet–Linevo paleodepression, Umet–Linevo sequence, Russian platform, palynozone, spores, conodonts.

In 1998, the Commission on Devonian Stratigraphy of the Russian Interdepartmental Stratigraphic Committee introduced the Volgograd Horizon into the Devonian stratigraphic scheme for the Russian platform and adjacent areas, and placed it between the Livny Horizon of the Frasnian Stage and the Zadonsk Horizon of the Famennian Stage (*Postanovleniya...*, 1999). The horizon corresponds to a hiatus recorded in the most part of the Russian platform, its central areas included. Though paleontological characteristics of the horizon stratotype are known, its bed-by-bed description has not been published so far. Such a description is necessary for the Volgograd Horizon to be treated as a valid stratigraphic unit. This article is to fill the gap.

The Frasnian–Famennian boundary sequences of a considerable thickness were first recovered by drilling in the 1960s in the Volgograd (boreholes 30-Linevo and 30-Umet) and Saratov (Borehole 3-Ilovlya) areas (Karpov *et al.*, 1960). The boreholes are concentrated in the Umet–Linevo paleodepression occupying the northeastern Volgograd and southern Saratov areas (Fig. 1). Between the Livny Horizon of the Frasnian Stage and the Zadonsk Horizon of the Famennian Stage, there have been recovered the carbonate–terrigenous deposits, which were divided into the lower Linevo sequence of the upper Frasnian Substage and the upper Umet sequence of the lower Famennian Substage according to their lithological features and fossil content (Karpov *et al.*, 1962; Karpov, 1970). Brachiopods from both units were described by Shevchenko, ostracodes were studied by Nechaeva (Karpov *et al.*, 1962) and spores by Nazarenko (Chibrikova and Nazarenko, 1974).

In the early Famennian, the greater part of the Russian platform, the Lower Volga region included, experienced elevation that caused a break in sedimentation, as it is evidenced by sharp changes in lithological composition and fossil assemblages, and by appearance of coarse-grained sandstones and limestone conglomerates at the base of the Zadonsk Horizon. However, the carbonate and clayey sedimentation continued in the Umet–Linevo paleodepression (Fig. 1).

In the early 1980s, the Linevo–Umet sequence was studied to use its thickness variations in a search of productive reef structures in the underlying upper Frasnian deposits. It was found out that the Linevo–Umet deposits had the noticeably reduced overall thickness, less than one meter in some places, above the organogenic structures (Figs. 1 and 6). Investigation of the beds penetrated in the Lomovo and Kotovo areas revealed a major change in palynological assemblages at the base of the Linevo sequence according to which it and, consequently, the entire Linevo–Umet section was attributed to the Famennian Stage (Mantsurova, 1987). Subsequent restudy of collected fossils confirmed the conclusion (Dan'shina *et al.*, 1995). Usually, the Linevo sequence that accumulated in the shallow shelf settings contains the reworked faunal remains of Livny age, which gave rise to the previous incorrect conclusion about the Frasnian range of the unit (Karpov *et al.*, 1962; Karpov, 1970). Because of facies peculiarities and scarcity of faunal remains, the Linevo–Umet biostratigraphy is mainly based on palynological data. The further analysis and correlation of East European Devonian palynological assemblages (Fig. 9) were used to establish the Devonian spore zonation in the region. The Linevo–Umet deposits and their age analogs (Fig. 8) were referred to the *Corbulispora viminea–Geminospora vasjamica* Zone and distinguished as the

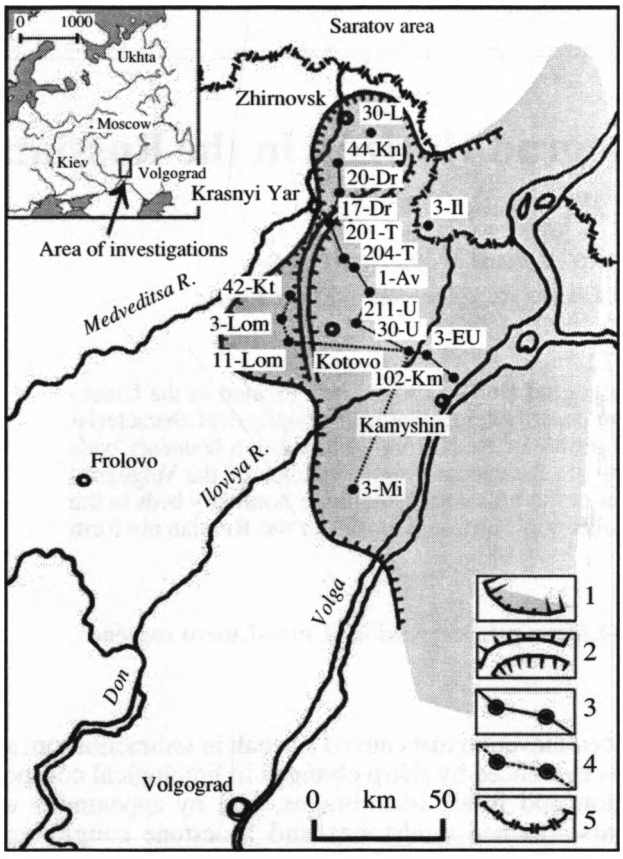


Fig. 1. Localities of borehole sections: (1) distribution limits of the Volgograd Horizon; (2) discontinuous reef zone presumably lacking the Volgograd deposits above the upper Frasnian organogenic buildups; (3) boreholes and correlation profiles; (4) boreholes and geological profile; (5) administrative boundary of the Volgograd area. Boreholes: (30-L) 30-Linevo, (44-Kn) 44-Krasnyi Yar, (20-Dr) 20-Doroshevo, (17-Dr) 17-Doroshevo, (3-Il) 3-Ilovlya, (201-T) 201-Tarasovka, (204-T) 204-Tarasovka, (42-Kt) 42-Kotovo, (11-Lom) 11-Lomovo, (3-Lom) 3-Lomovo), (1-Av) 1-Avilovo, (30-U) 30-Umet, (211-U) 211-Umet, (3-EU) 3-Eastern Umet, (102-Km) 102-Kamyshin, (3-Mi) 3-Mirnaya.

regional Volgograd Horizon of the Famennian Stage (Nazarenko *et al.*, 1993; Avkhimovitch *et al.*, 1993; Mantsurova and Tsygankova, 1995).

Below, we present the bed-by-bed description of the Volgograd Horizon sections of the stratotype area, the Medveditsa–Volga interfluve of the Umet–Linevo paleodepression of the northeastern Volgograd area (Fig. 1).

Each of the borehole sections under description is of particular significance. The borehole sections 44-Krasnyi Yar and 3-Lomovo compose the Volgograd Horizon stratotype. The section of Borehole 11-Lomovo is proposed for the parastratotype that yielded conodonts characteristic of the middle subzone of the *Palmatolepis triangularis* Zone.

Section of the Borehole 44 of the Krasnyi Yar area. The Volgograd Horizon was penetrated in 1966 by Borehole 44 in the Krasnyi Yar area, near the Krasnyi Yar Village of the Volgograd area (the Linevo paleodepression, Figs. 1 and 2). Its beds succession is described below from the top downward. The core samples were described by A.E. Pominova and N.V. Solov'eva. The middle and lower parts of the Volgograd Horizon were penetrated in the depth interval of 2242–2410 m: The core recovery was 31%.

Depth interval of 2296–2300 m (core recovery 2.2 m): dark gray, pelitic, calcareous mudstone (Sample 1); an admixture (3–7%) of quartz grains of fine silt size, fine organogenic detritus (3–5%), spores, and coalified plant detritus are identified in thin section. The rock is slightly dolomitized, showing horizontal bedding that is emphasized by orientation of inclusions.

Depth interval of 2300–2304 m (core recovery 2.8 m): dark gray mudstone (Sample 2) similar to the previous one; quartz grains (3–5%), which are scattered or accumulated in minute lenses, are visible in thin section. The mudstone contains calcitic sponge spicules, fragmentary algae (*Girvanella*), coalified plant detritus, and large spores. The bedding is horizontal, locally lenticular.

Depth interval of 2304–2308 m (core recovery 3.15 m): dark gray, highly clayey marl grading into mudstone (Sample 3); thin sections show silt admixture (1 or 2%), dolomitization (3%), and pyritization (2 to 3%). Mudstone contain ostracode shells with crystalline casts, calcitic sponge spicules, frequently pyritized organogenic detritus (5–7%), small spores, and fragments of plant tissue. Thin horizontal bedding reflects alternation of clayey and silty laminae.

Fine alternation of marl (Sample 4) and clastic micrite with algal fragments (*Girvanella*), calcitic sponge spicules, and ostracode shells is characteristic of the lower bed part.

Depth interval of 2308–2313 m (core recovery 1.7 m): fine interbedding of dark gray clayey marl similar to that from the interval of 2304–2308 (Sample 4) and light gray micrite (Sample 5); marl laminae vary in thickness from 0.2 or 0.3 to 2.0 cm and those of limestone, from 0.2 or 0.3 to 5.0 cm. Bedding is horizontal.

Depth interval of 2313–2317 m (core recovery 1.2 m): light gray, probably algal micrite (Sample 6) with minor (1 or 2%) content of quartz grains of fine silt size; rocks enclose single ostracode shells, minor amount of coalified detritus, and small spores. Usually they are massive, though open pores (3 or 4%) with crustified margins of calcite crystals can be observed in some places.

Depth interval of 2317–2321 m (core recovery 0.75 m): interbedding of clayey limestones (Sample 7) and calcareous mudstone (Sample 8); in thin section, limestone (Sample 7) interbeds show presence of calcitic sponge spicules in variable amounts and bear admixture of quartz grains (2 to 3%) of fine silt size and single ostracode shells.

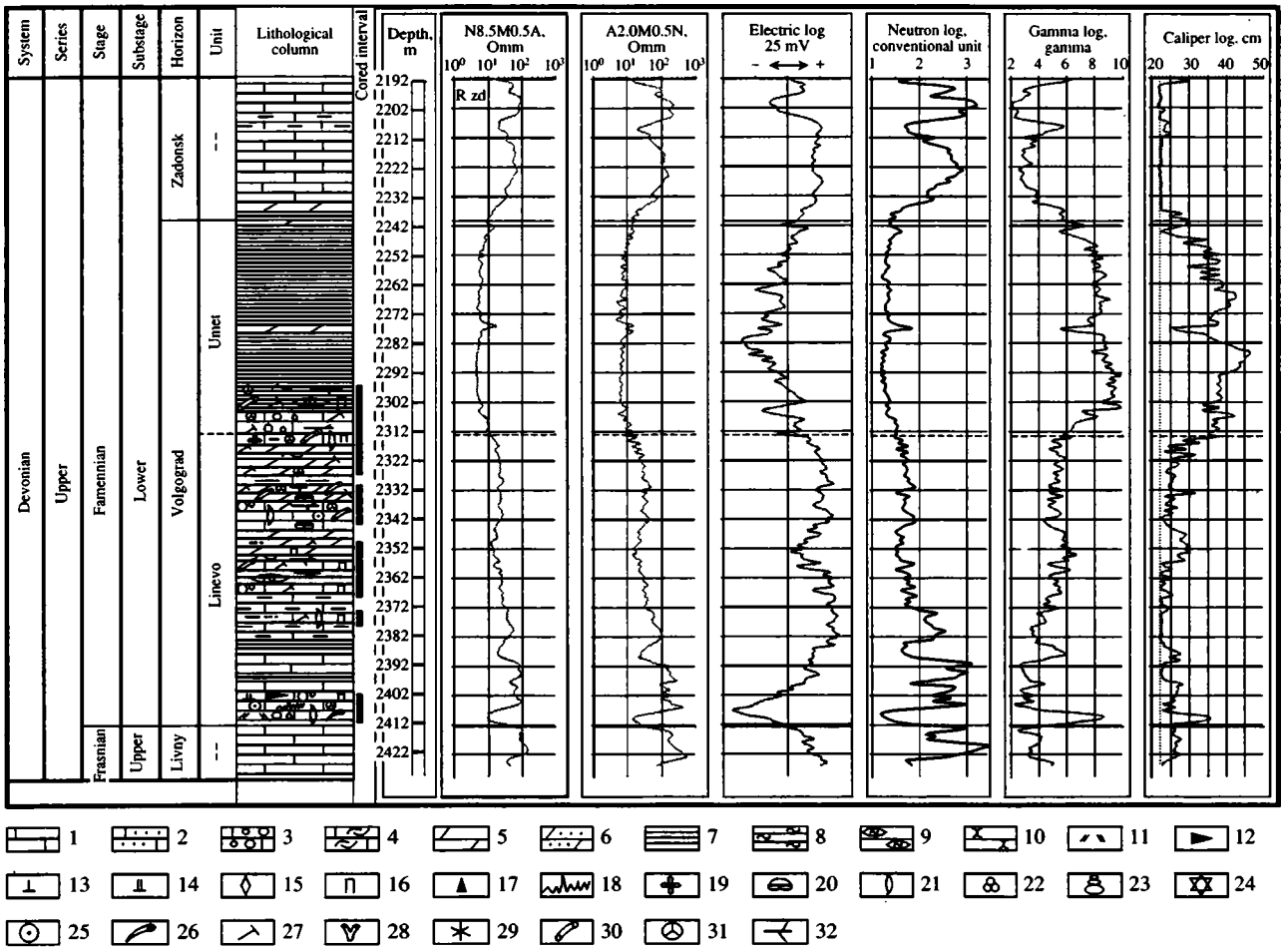


Fig. 2. Lithology and logging data for the Volgograd Horizon stratotype penetrated by Borehole 44-Krasnyi Yar in the Volgograd area: (1) limestone; (2) silty; (3) clastic, and (4) organogenic-detrital limestone; (5) marl; (6) silty marl; (7) mudstone; (8) silty mudstone; (9) siltstone lenses; (10) sandstone; (11) tuffaceous material; (12) bitumenization; (13) calcitization; (14) dolomitization; (15) recrystallization; (16) pyritization; (17) silicification; (18) stylolites; (19) cavernous rock; (20) brachiopods; (21) ostracodes; (22) foraminifers; (23) gastropods; (24) echinoderms; (25) crinoid columnals; (26) solitary corals; (27) sponge spicules; (28) bryozoans; (29) algae; (30) *Girvanella*; (31) plant spores; (32) plant detritus.

Depth interval of 2321–2323 m (core recovery 1.55 m): highly calcareous marl (Sample 9) similar in thin section to rocks from the depth interval of 2317–2321 m.

Depth interval of 2323–2326 m (core recovery 0.7 m): highly clayey marl (Sample 10) grading into mudstone in some interbeds; thin sections shows that marl interbeds are dolomitized and bear pyrite concretions, single calcitic ooliths, and small calcitic sponge spicules.

Depth interval of 2329–2333 m (core recovery 1.0 m): clayey marl-like limestone (Sample 11); thin sections show alternation of clayey micrite laminae bearing either the calcitic sponge spicules, or algae (*Girvanella*). The deposits showing horizontal bedding are pyritized (1 or 2%).

Depth of 2333–2338 m (core recovery 1.8 m): clayey marl-like limestone (Sample 12) with mudstone interbeds containing brachiopod shells; under microscope, the rock is similar to that from the preceding depth interval.

Depth interval of 2338–2340 m (core recovery 1.5 m): gray fine-grained to micritic, highly clayey limestone (Sample 13) with interbeds (up to 5.0 cm thick) of dark gray thin-laminated strongly calcareous mudstone (Sample 14); limestones contain fragments of brachiopod shells. In thin section, limestones are clotted, organogenic. They include round clasts (2–3 mm) of dark gray clayey organogenic–detrital, algal–ostracode limestones with pellets. Organic remains are represented by crinoid columnals, ostracode shells, small pelecypods, single brachiopod shells, calcitic sponge spicules, spheroids, and fragments of algae (*Girvanella*). Horizontal wavy bedding is shown by orientation of organic and pyritized plant remains, and by presence of clayey–bituminous laminae.

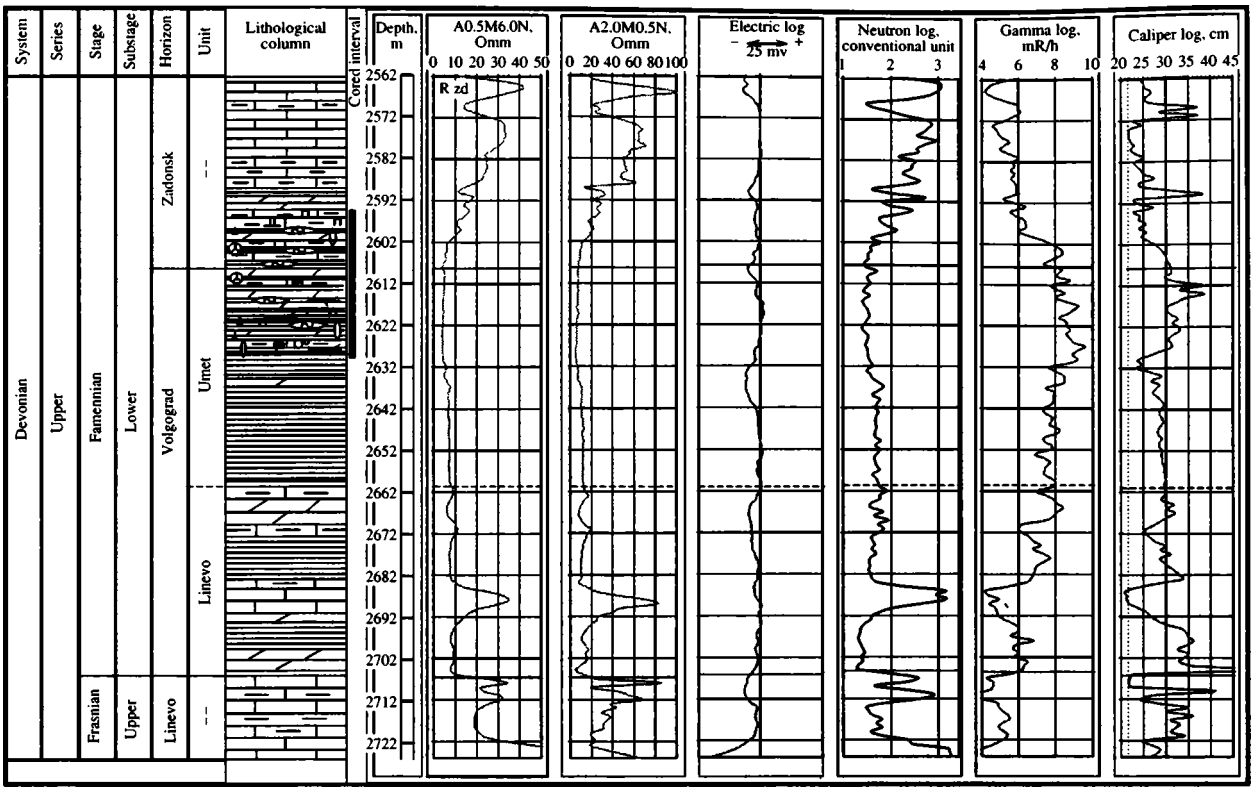


Fig. 3. Lithology and logging data for the Volgograd Horizon stratotype penetrated by Borehole 3-Lomovo in the Volgograd area (symbols as in Fig. 2).

Depth interval of 2340–2343 m (core recovery 1.6 m): dark gray, highly clayey, compact limestone (Sample 14a).

Depth interval of 2348–2353 m (core recovery 2.6 m): dark gray, highly clayey, dense marls (Sample 15).

Depth interval of 2353–2358 m (core recovery 2.8 m): strongly calcareous dense mudstone (Sample 16) of dark to greenish gray color with interbeds (2–3 cm) of light gray micrite (Sample 17). Under microscope, mudstones (Sample 16) are dolomitized, dense, pyritized, showing admixture (2–3%) of quartz grains of fine silt size, presence of coalified detritus, and accumulations of alga clots (*Girvanella*) 0.2 to 1.0 mm in size. The clayey limestone (Sample 17) with clasts of algal limestones contains abundant calcitic sponge spicules, and single ostracode shells.

Depth interval of 2358–2363 m (core recovery 1.0 m): dark gray silty mudstone (Sample 18) grading into siltstones in some interbeds; rocks reveal lenticular bedding with elements of convolute texture and worm trails. Thin section (Sample 18) shows that silty mudstone encloses siltstone lenses composed of quartz grains of fine silt size and plates of mica (muscovite), as well as pyrite concretions and fragments of plant tissue.

Depth interval of 2363–2369 m (core recovery 2.0 m): dark gray, highly clayey micrite (Samples 19 and 20) enclosing interbeds of dark gray mudstone (Sample 21). Under microscope, limestones are similar to those from the interval of 2353–2358 m. Mudstones contain silty admixture (5–8%).

Depth interval of 2373–2378 m (core recovery 2.5 m): highly clayey dense micrite of dark to brownish gray color (Sample 22). Under microscope, limestone shows admixture (2–3%) of quartz grains of fine silt size and presence of calcitic sponge spicules and single ostracode shells with crystalline casts. Fine horizontal bedding is emphasized by orientation of long axes of sponge spicules.

Depth interval of 2400–2405 m (core recovery 1.9 m): dark gray, highly clayey, irregularly dolomitized micrite (Sample 23) with interbeds of slightly clayey limestone (Sample 24). Under microscope, limestones (Samples 23 and 24) are bituminous (10–12%), having dolomitization spots (20%), and containing unsorted clasts (0.1 to 3–5 mm) of algal limestones, quartz grains, crinoid columnals, and ostracode shells. Rocks are pyritized (5%), having stylolites.

Depth interval of 2405–2410 m (core recovery 0.5 m): light gray, irregularly dolomitized, very dense limestone (Sample 25). In thin section, the rock is micritic, showing dolomitization and kaolinization spots, presence of organogenic limestone clasts, noticeable admixture (20%) of crystalline clasts (fragments of

bipyrimidal quartz crystals, zoned plagioclase, big biotite scales, greenish chloritoid, and others). It yields ostracode shells, crinoid columnals, and single coral fragments.

According to logging data, Borehole 44-Krasnyi Yar penetrated mudstones of high gamma-activity and low electric resistance at the base of the Volgograd Horizon (depth interval of 2407–2412 m); the hole diameter is increased within this interval. Below, at the top of the Livny Horizon (depth interval of 2412–2422 m), there are fissured, cavernous, dolomitized, clayey (5–14%) and porous (4–11%) limestones.

Section of Borehole 3 of the Lomovo area. The upper part of the Volgograd Horizon (the Umet Unit) and the Volgograd–Zadonsk boundary beds were penetrated in the interval of 2595–2629 m by borehole drilled in 1979 near the town of Kotovo of the Volgograd area (Figs. 1 and 3). The core samples were described by L.A. Klimova and N.V. Dan'shina. The Volgograd Horizon 100 m thick was penetrated in the depth interval of 2606–2706 m. In the cored interval (2606–2629 m), core recovery is 54%. Core was also obtained from the Volgograd–Zadonsk boundary beds. The Zadonsk Horizon was penetrated in the depth interval of 2595–2606 m. The sampled deposits are described below.

The Zadonsk Horizon

Depth interval of 2595–2600 m (core recovery 0.7 m): dark gray micrite (Samples 1 and 3) with threadlike clayey laminae and interbeds of dark gray to black micritic marls (up to 4 m thick). In the clayey limestones, there are nest-shaped inclusions (up to 0.8 cm in diameter) of pyrite and dolomite, vertical fissures every 2–3 cm, and worm trails. Rocks are horizontally bedded. Under microscope, marls (Sample 2) consist of clayey (50%) and carbonate (40%) matter that is dolomitized (2–3%) and pyritized (1–2%). In thin section, limestone (Sample 3) is cryptograined, having granular pyritized inclusions (0.2–0.3 mm) of recrystallized organogenic detritus. Rare stylolites are dolomitized.

Depth interval of 2600–2606 m (core recovery 0.89 m): interbedding of dominating mudstone (Sample 5) with marl (Sample 6), and limestone (Sample 7); laminae of 2.5, 3.0 and 0.7 cm thick, respectively. The deposits are dark gray, variably calcareous, with abundant ostracode and rare pelecypod shells on the bedding planes. In thin section, the mudstone (Sample 5) is calcareous (21–22%), dolomitized (6.34% according to the chemical analysis), pyritized (2%), and having flaky accumulations of leucoxene and plant spores. Lenticular–horizontal bedding reflects the presence of lenses and discontinuous laminae of coarse-grained quartz siltstone with single grains of epidote and zircon, and with muscovite scales, all cemented by authigenic calcite and pyrite. Cement is of the basal type. Under microscope, marl (Sample 6) is composed of clayey (40%) and carbonate (45.35% according to chemical analysis) matter that is dolomitized (5.65%) and pyritized (3–5%) and contains rare fragments of small unidentifiable shells and scarce spores. The horizontal bedding is outlined by interbeds with quartz grains of silt size, abundant (10–15%) muscovite scales, coalified plant detritus, and recrystallized organogenic remains. In thin section, limestone (Sample 7) appears to be micritic, dolomitized, and locally as clayey as marl. Dolomite rhombs (0.01–0.1 mm) are regularly dispersed through the thin section. There are also aggregates of crystals (0.05–0.1 mm), and some rhombs are of a zoned structure. Pores in the rocks (7–10%) are rounded and isometric (0.03–0.1 mm). Multidirectional fissures represent another characteristic feature.

The Volgograd Horizon

Depth interval of 2606–2611 m (core recovery 4.0 m): alternating laminae of mudstone (10–20 cm) and limestone (1.0–3.0 cm). Mudstone (Sample 8) is dark to greenish gray, irregularly calcareous, poorly pyritized, flaggy to foliate, having ostracode shells and coalified plant detritus on the bedding planes. Thin section shows that mudstone is slightly dolomitized, bearing admixture of silty quartz grains and fine organogenic detritus. The gray clayey micritic limestones (Samples 9 and 10) reveal short wedge-shaped fissures (0.1 cm wide) filled with calcite. Under microscope, these dense, slightly pyritized limestones show some lenses containing quartz grains, shells of ostracodes and small pelecypods.

Depth interval of 2611–2618 m (core recovery 3.8 m): interbedding of marl, limestone, and subordinate mudstone (Sample 19). Limestones interbeds yield well-preserved brachiopod shells. Thin sections of marls (Samples 16 and 17) show them to be highly clayey, grading into mudstone, poorly pyritized, and having lenses of silty quartz grains associated with feldspar grains, rare big scales (up to 0.2 mm) of chloritized biotite, and scattered coalified plant detritus. The lenses are cemented by calcite and, locally, by kaolin. Their cement is of the basal type. In thin sections, micritic limestones (Samples 18 and 20) are dense, poorly pyritized (1–2%), and having admixture (1–2%) of quartz grains, colorless mica scales, and coalified detritus.

Depth interval of 2618–2623 m (core recovery 2.1 m): dark to greenish gray mudstone (Sample 21). Some laminae of the rock (1.5–2.0 cm) are silty, irregularly calcareous up to the state of marl (Sample 22). The rock contains rare brachiopod shells. Thin sections show that calcareous and pyritized (5–7%) mudstones enclose quartz siltstone laminae, chloritoid grains, coalified plant detritus, and spores. Bedding of the rocks is horizontal.

Depth interval of 2623–2629 m (core recovery 2.5 m): dark and greenish gray mudstones (Samples 24 and

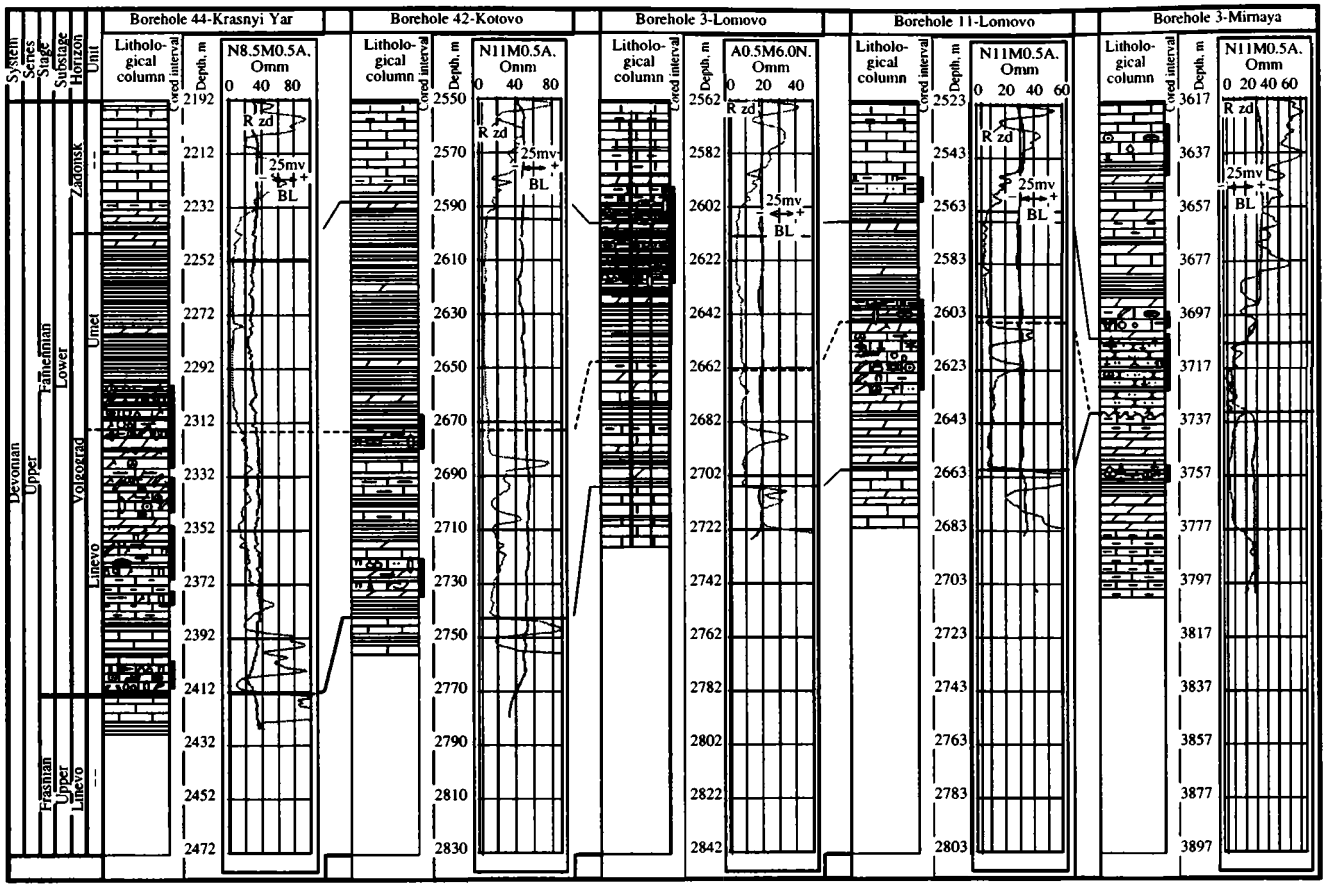


Fig. 4. Correlation of Frasnian–Famennian boundary beds of the Umet–Linevo paleodepression, the Volgograd area (symbols as in Fig. 2).

25); rocks are irregularly calcareous, locally up to the state of marl. On their bedding planes, there are ostracode and brachiopod shells. Mudstones enclose rare and thin (up to 3.0 cm) laminae of gray clayey limestone (Sample 26). In thin sections, they reveal calcareous and silty admixture. Thin lenses and laminae of quartz siltstones occurring in the rocks are responsible for their lenticular–horizontal bedding. The rocks are pyritized, slightly dolomitized, containing coalified plant detritus, spores, and unidentifiable fine organogenic detritus. In thin sections, the micritic dense limestone shows clasts of organogenic limestones, admixture of silty quartz grains, shells of small pelecypods and ostracodes, remains of unilocular foraminifers, brachiopods, echinoderm plates, and algae (*Girvanella*).

According to logging data, the base of the Volgograd Horizon (depth interval 2692–2706 m) penetrated by Borehole-3 in the Lomovo area corresponds to fissured and cavernous clayey (15–23%) limestones characterized by low electric resistance; the hole diameter is increased within this interval. Below, at the top of the Livny Horizon (interval 2706–2713 m), there are dense limestones, the middle part of which includes a mudstone bed nearly one meter thick. Clay content in limestones having high electric resistance is less than 5%; hole diameter in them is nominal.

Section of Borehole 11 of the Lomovo area. The Volgograd Horizon was penetrated by Borehole 11-Lomovo (depth interval 2560–2660 m) drilled in 1980 near the town of Kotovo of the Volgograd area (Figs. 1 and 4). The core samples were described by N.V. Dan'shina and T.I. Fedorenko. The horizon is 100 m thick (core recovery 20%). The cored intervals correspond to the lower part of the Volgograd (depth interval of 2597–2630 m) and Zadonsk (depth interval of 2551–2560 m) horizons, the bed-by bed description of which is presented below.

The Zadonsk Horizon

Depth interval of 2551–2560 m (core recovery 3.08 m): alternating limestone (Samples 4, 6, and 7) and mudstone (Sample 5) beds. Limestones are dark gray, micritic, and sometimes as clayey as marl. They show pyritized worm trails and well-preserved brachiopod shells. Mudstones are dark gray and irregularly calcareous. In thin section, they appear to be dolomitized, pyritized (5–7%), bituminous, and enclosing siltstone

lenses and clasts of algal limestones. Carbonate cement of siltstones is of the basal type. Horizontal bedding is outlined by alternating limestone (6–10 cm) and mudstone (0.1–1.0 cm) laminae and by parastylolites. Fissures, predominantly vertical ones, are characteristic of the rocks.

The Volgograd Horizon

Depth interval of 2597–2605.5 m, Bed 1 (core recovery 0.4 m): dark gray, irregularly clayey, micritic limestones (Sample 14) with vertical fissures (every 3 cm), rare pyrite nests, and worm trails. The rock is composed of clayey (20%) and carbonate (70%) matter. Embryonic stylolites of clayey–bituminous matter can be observed.

Bed 2 (core recovery 0.6 m): dark to greenish gray mudstones (Sample 16). These flaggy micaceous can be as calcareous as marl. They enclose lenticular laminae of gray pelitomorphic limestones (Sample 17) with rare brachiopod shells. On the bedding planes, there are ostracode shells.

Depth interval of 2605.5–2612 m, Bed 1 (core recovery 2.35 m): fine interlamination of limestones (Sample 18) and mudstones (Sample 20). The laminae have gradational boundaries. Deposits are dark gray, highly clayey, irregularly calcareous, enclosing lenticular laminae (1.5 cm) of hard gray micritic limestone (Sample 19).

Bed 2 (core recovery 2.3 m): marl intercalations in prevalent mudstones. Marl (Sample 23) is dark to greenish gray, micritic, containing brachiopod shells. Mudstone (Sample 21) is dark gray, flaggy, locally as calcareous as marl (Sample 22). On bedding planes, there are mica accumulations and coalified plant detritus. In thin sections, mudstones are silty, enclosing indistinct lenses of silt-sized quartz grains, ostracode shells, bryozoan fragments, and algae remains (*Girvanella*), all cemented with authigenic calcite (cement of the basal type). Marl (Sample 22) is silty (15%) and pyritized (10%), bearing pyrite concretions (0.2 cm across). Marl yields calcitic sponge spicules (5–7%), recrystallized styliolines, spores (2 or 3%), and fragmental plant tissue of reddish brown color.

Bed 3 (core recovery 1.75 m): gray to brownish gray limestones (Samples 24–27), which are, micritic, irregularly clayey, dolomitized, recrystallized, hard, and massive. Because of multidirectional sutures, they resemble conglomerate. Curved vertical fissures up to 2 mm wide are filled with calcite. There are inclusions (5.0 × 2.0 cm) of calcite in association with sulfates and pyrite. Pyrite crystals occur in aggregates (1.5 cm). Abundant brachiopod shells were found in the upper (Sample 24) and lower (Sample 26) parts of the bed. Crystals of calcite and anhydrite fill inner cavities of some shells. Microscopic study shows that limestones (Sample 25) are clotted, algal (*Girvanella, Rothpletzella*), porous to cavernous (10–12%), having inclusions of authigenic quartz crystals. Caverns are filled with calcite, and in places with dolomite and sulfate. Crystals of dolomite and anhydrite are big (0.15–0.20 mm). Pore margins are crustified with small calcite crystals. Single calcitic ooliths and large gastropod shells are also distinguished.

Depth interval of 2612–2622.7 m, Bed 1 (core recovery 0.25 m): gray, locally brownish limestone (Sample 28) of conglomerate-like appearance due to multidirectional sutures. They are similar in general to the limestones of Bed 3 from the depth interval of 2605–2612 m.

Bed 2 (core recovery 0.55 m): gray to greenish gray organogenic-detrital limestone (Sample 29); rock is dolomitized and irregularly clayey. Coquina interbeds contain abundant brachiopod and pelecypod shells. Discontinuous clayey laminae outline indistinct bedding. Unfilled joined fissures are characteristic of the rock. Under microscope, organogenic detritus is represented by fragments of brachiopods, pelecypods, gastropods, rare crinoid columnals, echinoderm plates, and algae (*Girvanella*). Clasts coated with algae and pellets were also found in micritic limestone. Clayey dolomitized matrix of limestones is irregularly crystalline, impregnated with dark brown bituminoid (filling interstices).

Bed 3 (core recovery 1.45 m): gray (Sample 30) to dark gray (Sample 31) micritic limestones, clayey varieties of which are dolomitized. On bedding planes, there are casts and valves of brachiopod and pelecypod shells and crinoid columnals. Inner cavities of some brachiopod shells are filled with calcite crystals, which are yellowish due to bitumen impregnation. Rocks reveal the lenticular bedding. Limestones show differently oriented, joined and open wedge-like fissures (up to 4.0 cm) filled with calcite.

Bed 4 (core recovery 0.20 m): dark to greenish gray mudstone (Sample 32); rocks are calcareous, flaggy to foliate, poorly micaceous, pyritized, bearing coalified plant detritus on bedding planes. Under microscope, the carbonate–hydromica rock matrix encloses small quartz grains, fine scales of mica (muscovite), small ostracode shells, balls and threads of algae (*Girvanella*), and fine coalified plant detritus. Bedding is horizontal.

Bed 5 (core recovery 1.20 m): dark gray, organogenic–detrital, irregularly clayey limestones (Samples 33 and 34) with micrite clasts and algal concretions. The upper part includes coquina interbeds (up to 6 cm) composed of brachiopod shells and crinoid columnals. Rocks are pyritized. Pyrite occurring in form of separate inclusions (up to 1.2 cm) is in paragenetic sequence with calcite and dolomite. Conglomerate-like texture is due to irregular distribution of organogenic matter and sutures of different orientation. Joined subvertical fissures are widespread.

Bed 6 (core recovery 0.45 m): dark gray, clayey, clotted limestones (Samples 35 and 36) with calcitic

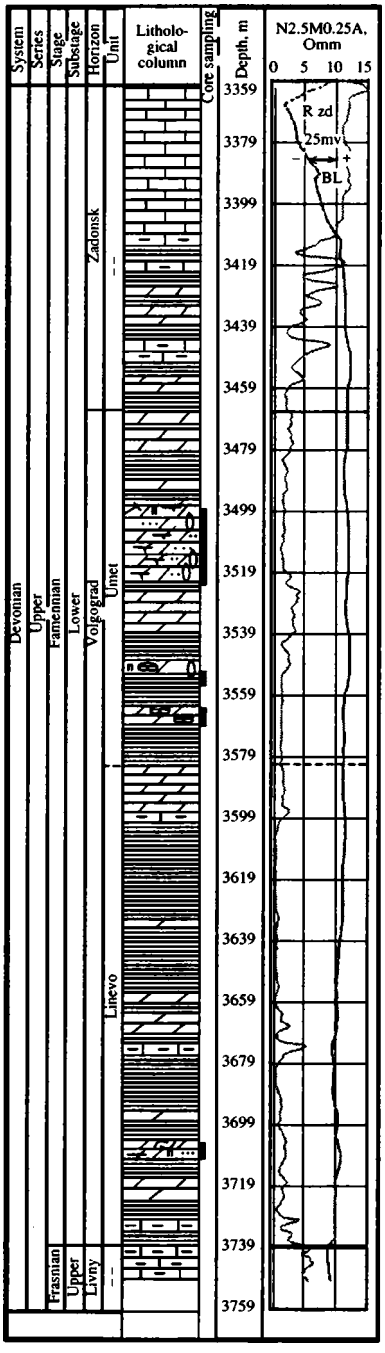


Fig. 5. Lithology and logging data for the Volgograd Horizon penetrated by Borehole 30 in the Umet area (symbols as in Fig. 2).

ooliths in the upper part of the bed and with abundant crinoid columnals and well-preserved brachiopod shells in the lower part. Rocks are hard, having single caverns (1 × 2 cm) partly filled with authigenic coarsely crystalline calcite. These massive limestones are intensively fissured. Fissures are subvertical and open.

Bed 7 (core recovery 0.40 m): are dark to greenish gray laminated marls (Samples 37 and 38). Rocks enclose interbeds of dark gray, organogenic–detrital limestones (Sample 39) in the upper part of the bed, which are similar to those from Bed 2 (Sample 29, depth interval 2612–2622 m). Under microscope, marls are bituminous, with silty quartz grains (7–15%), thin scales of mica (muscovite), small concretions of globular pyrite (5%), abundant balls and separate tubes of algae (*Girvanella*), small ostracode shells, and coalified plant detritus (5–7%). Fine horizontal lamination is outlined by orientation of the inclusions along the bedding planes. Limestone (Sample 39) is organogenic–detrital (clasts of foraminifers and ostracodes) and irregularly clayey; most clayey spots are dolomitized. It is pyritized, irregularly crystallized, and showing stylolites. Yellow–brown bitumen stains are confined to stylolites and spots with largest calcite crystals. Non-carbonate fraction (2–3%) consists of authigenic idiomorphic crystals of quartz.

Depth interval of 2622–2630 m (core recovery 2.7 m): alternating mudstones, marls and limestones, which grade into each other. Horizontal bedding is indistinct. Rocks are fissured. Fissures are open, subvertical, partly intersecting. Dark gray, clotted limestones with abundant ostracodes, brachiopod shells, and rare corals (Samples 40–43, 48, and 49) show micritic laminae and dolomitized spots (Sample 43). Under microscope, Samples 41 and 43 are classed with micritic limestones, irregularly clayey, silty (2–3%), bituminous, dolomitized, and pyritized (small inclusions of globular pyrite). They contain small thin shells of ostracodes, unilocular foraminifers, coalified plant detritus, and single tubes of algae (*Girvanella*). Mudstones (Samples 44 and 46) are dark gray, almost black in some interbeds, locally strongly calcareous, micaceous, bearing coalified plant detritus on bedding planes. In the upper part of the bed, mudstones (Sample 44) contain lenses (up to 1.5 cm) of dolomitized limestones (Sample 45). Under microscope, mudstones (Samples 44 and 46) are strongly calcareous, silty (3–5%), having small spores and bearing abundant fragments of yellow–brown plant tissue, rare calcitic sponge spicules, single foraminiferal tests, and echinoderm plates. Rocks are pyritized, locally dolomitized, showing single glauconite grains. The dark gray marl (Sample 47) is micaceous. Microscopic examination shows that marls, like mudstones, contain silty quartz grains (5–7%), fragments of plant tissue, large scales of colorless mica, isolated tubes of algae (*Girvanella*), and calcitic sponge spicules.

According to logging data, the base of the Volgograd Horizon penetrated by Borehole 11 in the Lomovo area (depth interval of 2653–2660 m) corresponds to marls overlying a mudstone bed 1.7 m thick. Marls exhibit high gamma-activity and low electric resistance; hole diameter is increased within the indicated interval. Underlying limestones (the top of the Livny Horizon, depth interval of 2660–2667 m) are dense, somewhat porous (up to 10%) in the upper part, clayey

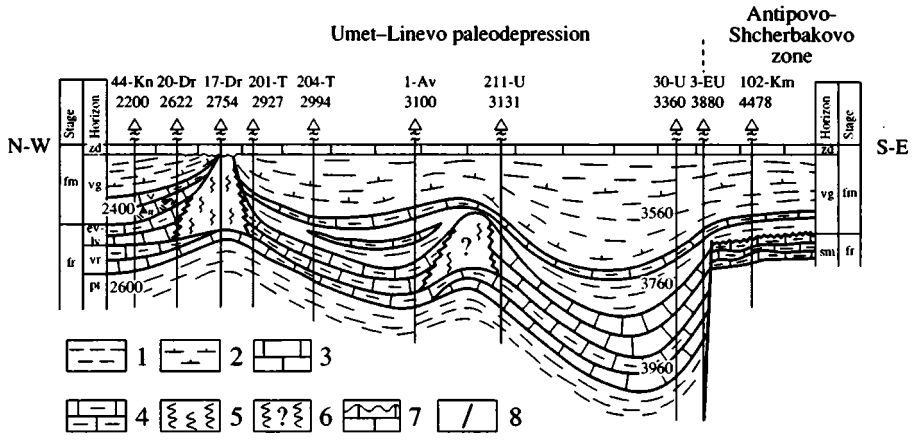


Fig. 6. Paleogeological profile of Frasnian and Famennian deposits in the Umet–Linevo paleodepression of the Volga region: (1) mudstone; (2) calcareous mudstone; (3) limestone; (4) clayey dolomitized mudstone or marl; (5) reefal limestone; (6) presumably reefal buildups; (7) erosional surface; (8) fault. Stratigraphic units: (fr) Frasnian Stage; (sm) Semiluki Horizon; (pt) Petino Horizon; (vr) Voronezh Horizon; (ev–lv) Evlanovo and Livny horizons; (fm) Famennian Stage; (vg) Volgograd Horizon; (zd) Zadonsk Horizon (other symbols as in Fig. 2).

(less than 5%). In the upper part, their electric resistance is high, and hole diameter remains unchanged.

The structure of the Volgograd Horizon and its stratigraphic relationships with the underlying and overlying horizons are shown in the schemes and paleogeological profile (Figs. 4, 5, and 6). In the Umet–Linevo paleodepression, the Volgograd Horizon rests on the Livny Horizon and thins out westward. To the east, toward the North Caspian depression, it locally overlies with a hiatus the erosional surface of the Semiluki Horizon. The hiatus corresponds to the pre-Famennian sedimentation break in some sections of the old Antipovo–Shcherbakovo uplift (Fig. 6, Mantsurova, 2000).

The Volgograd Horizon conformably rests on the Livny beds (the *Grandispora subsuta* Subzone) represented by clayey limestones with interbeds of marls and mudstones in the relatively deep-water shelf sections and by biohermal limestones in the shallow-water shelf sections at the western edge of the Umet–Linevo paleodepression. In the former sections (Borehole 30-Umet), the Volgograd Horizon is largely composed of mudstones and marls with threadlike thin (to 3.0 cm) interbeds of gray pelitomorphic limestones (Fig. 5). The entire horizon is characterized by horizontal bedding. The mudstones are calcareous, pyritized (1–5%), with admixture of silty quartz grains (2–10%). Their rock-forming minerals are hydromica (70–80%) and kaolinite (30–20%). The mudstones contain rare ostracodes, brachiopods, and fine coalified plant detritus. Marls having characteristic greenish tint are frequently dolomitized (2–18%). The maximal thickness of the Volgograd Horizon is 320 m (Borehole 10-Southern Umet, Tsygankova and Mantsurova, 1998).

In the shallow-water shelf sections, the lower part of the Volgograd Horizon (the Linevo sequence) is more calcareous. Limestones are organogenic-detrital, irregularly clayey, enclosing interbeds of marls and mudstones. They contain more frequent calcitic sponge spicules, abundant algae (*Girvanella*), and fragments of tabulates and stromatoporoids. According to data of chemical analysis, the lower part of the horizon is more calcareous (43%) than the upper one (36%). The upper part (the Umet sequence) is composed predominantly of mudstones and marls containing quartz grains of fine silt size, fine coalified plant detritus, spores, rare shells of brachiopods and ostracodes.

The Livny–Volgograd boundary beds were penetrated in boreholes 40 (depth interval 2631–2636 m) and 43 (depth interval 2645–2652 m) of the Kotovo area (shallow-water shelf sections, Fig. 1). The Volgograd mudstones are thinner here and unconformably, with a sharp contact, overlie the Livny limestones. Numerous caverns and stylolites of the Livny organogenic-clastic limestones evidence for a minor hiatus. Clasts in the rocks are 0.1 to 5 cm in size, represented by irregularly clayey, dolomitized, clotted algal limestones. The hiatus may correspond to the lower subzone of the *Palmatolepis triangularis* Zone or to a part of it.

Basal beds of the Volgograd Horizon are silicified (Borehole 44-Kotovo, interval 2751–2759 m), locally bearing sulfates and pyrite (Borehole 43-Kotovo). Pyritization affected the reworked clasts of algal limestones and stromatoporoids of Livny age (Dan'shina *et al.*, 1995). In the section of Borehole 44-Krasnyi Yar (interval 2405–2410 m), these beds are composed of dense limestones with crystalloclastic admixture pointing to the volcanic activity of that time.

At the southwestern edge of the Umet–Linevo paleodepression, the carbonate–clayey deposits are substituted by terrigenous beds, because the Zadonsk basement projection was the main provenance of terrigenous material during the early Famennian time (Fig. 4). Sandstone beds 26 m thick were penetrated by Bore-

holes 2-Mirnaya (interval 3843–3847 m) and 3-Mirnaya (interval 3716–3726 m). These are medium-grained (0.2–0.35 mm), well sorted quartzitic sandstones bearing feldspar grains (microcline). Their porous–filmy cement consists of kaolinite. In places, the sandstones are quartzitic, showing the conformal type of cementation, and having inclusions of leucoxene and pyrite. These rocks are massive. Sandstones alternate with siltstones and calcareous mudstones in the upper part of the Volgograd Horizon.

The Volgograd–Zadonsk boundary beds were also penetrated in Borehole 3-Lomovo. The Volgograd mudstones (interval 2606–2611 m) are overlain here by the Zadonsk marls and limestones (interval 2595–2606 m). The limestones are micritic, clayey, slightly silty (2–5%), dolomitized, pyritized, bearing ostracode shells and plant spores. Their bedding is horizontal–lenticular. At the base of the Zadonsk Horizon, there are frequent interbeds of oolitic and organogenic-detrital limestones containing crinoid columnals, shells of ostracodes and brachiopods, and rare remains of bryozoans and echinoderms.

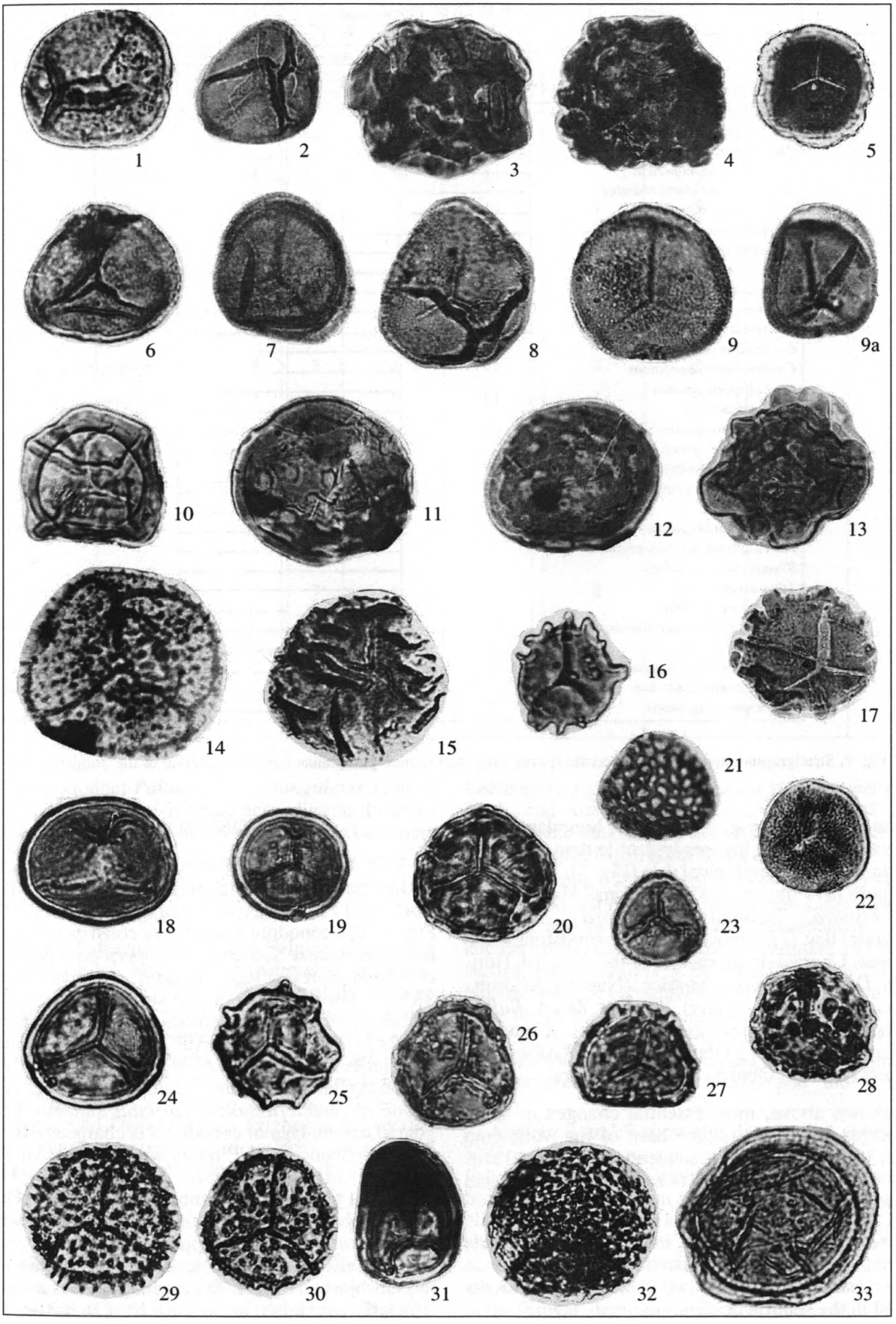
SPORES

Palynology of the Volgograd Horizon (the *Corbulispora viminea–Geminospora vasjamica* Zone) was studied in many borehole sections representing predominantly the shallow-water shelf deposits (Chibrikova and Nazarenko, 1974; Mantsurova, 1987; Dan'shina *et al.*, 1995; Mantsurova and Tsygankova, 1995). The spore assemblage from the horizon consists of 40–50 species, among which 18–24 species are most widespread (Fig. 7, Tables 1–3, and plate). It differs from the Livny spore assemblage (the *Grandispora subsuta* Subzone) in the complete absence of typical Frasnian species with filmy exine, such as *Grandispora subsuta* (Naz.) Obukh., *Auroraspora speciosa* (Naum.) Obukh., *Cristatisporites deliquescens* (Naum.) Arkhang., and *Membrabaculisporis radiatus* (Naum.) Arkhang., and also in appearance of *Cymbosporites boafeticus* (Tchibr.) Obukh., *Geminospora vasjamica* (Tchibr.) Obukh., *G. notata* (Naum.) Obukh. var. *microspinosus* Tchibr., *Corbulispora viminea* (Nekr.) Obukh., *C. semireticulata* (Tchibr.) Tchibr., *Reticulatisporites volgogradicus* Mants. sp. nov. msc., *Lophotriletes multiformis* Tchibr., and *Punctatisporites famenensis* (Naum.) Obukh. Dominant forms are *Converrucosisporites curvatus* (Naum.) Turnau, *Pustulatisporites pullus* (Naum.) Obukh., and *Stenozonotriletes definitus* (Naum. (Tables 1–3). In most sections, *P. pullus* (Naum.) Obukh. dominates in spectra characterizing the lower part of the horizon, whereas *C. curvatus* (Naum.) Turnau is a dominant form in the upper part. Frequently, abundance of *G. notata* (Naum.) Obukh. var. *microspinosus* Tchibr. increases from the base upward in the section.

Spores *Lophotriletes multiformis* Tchibr., *Cymbosporites boafeticus* (Tchibr.) Obukh., *Corbulispora semireticulata* (Tchibr.) Tchibr., and *Verrucosisporites evlanensis* (Naum.) Obukh. usually represent up to 5% of the spectra, but they do not occur in every section. The Volgograd deposits also contain *Retusotriletes pychovii* Naum., *Archaeozonotriletes variabilis* Naum., *Auroraspora varia* (Naum.) Ahmed., and *Reticulatisporites volgogradicus* Mants. sp. nov. Percentages of species most frequent in the Volgograd Horizon are presented in Table 3. Abundance of tetralete spores *Archaeozonotetraletes* Araslanova corresponds to 2 or 3% of spectra in the majority of sections.

Spores from the Famennian Volgograd Horizon of the Russian platform (the *Corbulispora viminea–Geminospora vasjamica* Palynozone) Magnification is 500).

(1, 2) *Geminospora notata* (Naum.) Obukh. var. *microspinous* Tchibr.: (1) Borehole 44-Krasnyi Yar, depth 2304–2307 m, Sample 2, (2) Borehole 42-Kotovo, depth 2675–2680 m, Sample 10; (3, 4) *Corbulispora viminea* (Nekr.) Obukh.: (3) Borehole 23-Romanovskaya, depth 3480–3487 m, Sample 10, (4) Borehole 4-Ostrovskaya, depth 2740–2746 m, Sample 4; (5) *Auroraspora varia* (Naum.) Ahmed. var. *minor* Naum., Borehole 42-Kotovo, depth 2671–2746 m, Sample 6; (6–9a) *Geminospora vasjamica* (Tchibr.) Obukh. et Nekr.: (6) Borehole 44-Krasnyr Yar, depth 2304–2308 m, Sample 6, (7) Borehole 23-Romanovskaya, depth 3480–3487 m, Sample 2, (8, 9) Borehole 3-Lomovo, depth 2606–2611 m, Sample 12, (9a) Borehole 11-Lomovo, depth 2605–2612 m, Sample 20; (10) *Knoxisporites dedaleus* (Naum.) Moreau-Benoit, Borehole 44-Krasnyi Yar, depth 2304–2308 m, Sample 2; (11, 12) *Dictyotriletes annae* Mants. sp. nov. msc., Borehole 3-Mirnaya, depth 3709–3716 m, Sample 41; (13) *Lophotriletes multiformis* Tchibr., Borehole 3-Lomovo, depth 2618–2623 m, Sample 21; (14) *Cymbosporites rovninae* Mants. sp. nov. msc., Borehole 97-Kamyshin, depth 4805–4810 m, Sample 10; (15) *Corbulispora semireticulata* (Tchibr.) Tchibr., Borehole 69-Novo-Korobkovo, depth 2633–2638 m, Sample 22; (16, 17) *Pustulatisporites pullus* (Naum.) Obukh.: (16) locality and sampling level as for fig. 2, (17) locality and sampling level as for fig. 5; (18) *Retusotriletes communis* Naum., locality and sampling level as for fig. 5; (19) *Retusotriletes pychovii* Naum., locality and sampling level as for fig. 5; (20) *Pustulatisporites famenensis* (Naum.) Obukh., locality and sampling level as for fig. 5; (21) *Reticulatisporites volgogradicus* Mants., sp. nov. msc., Borehole 3-Mirnaya, depth 3709–3716 m, Sample 41; (22, 23) *Punctatisporites famenensis* (Naum.) Obukh.: (22) locality and sampling level as for fig. 5, (23) Borehole 3-Lomovo, depth 2606–2611 m, Sample 12; (24) *Stenozonotriletes definitus* Naum., Borehole 14-Lomovo, depth 2670–2678 m, Sample 50; (25–27) *Converrucosisporites curvatus* (Naum.) Turn.: (25, 27) Borehole 44-Krasnyi Yar, depth 2304–2308 m, Sample 2, (26) locality and sampling level as for fig. 8, 9; (28) *Converrucosisporites curvatus* (Naum.) Turnau var. *medius* Kedo, locality and sampling level as for fig. 5; (29, 30) *Cymbosporites boafeticus* (Tchibr.) Obukh.: (29) Borehole 42-Kotovo, depth 2669–2671 m, Sample 1, (30) locality and sampling level as for fig. 15; (31) *Archaezonotriletes variabilis* Naum. var. *insignis* Senn., locality and sampling level as for fig. 5; (32) *Verrucosisporites evlanensis* (Naum.) Obukh., locality and sampling level as for fig. 5; (33) *Diaphanospora rugosa* (Naum.) Byvsch., locality and sampling level as for fig. 5.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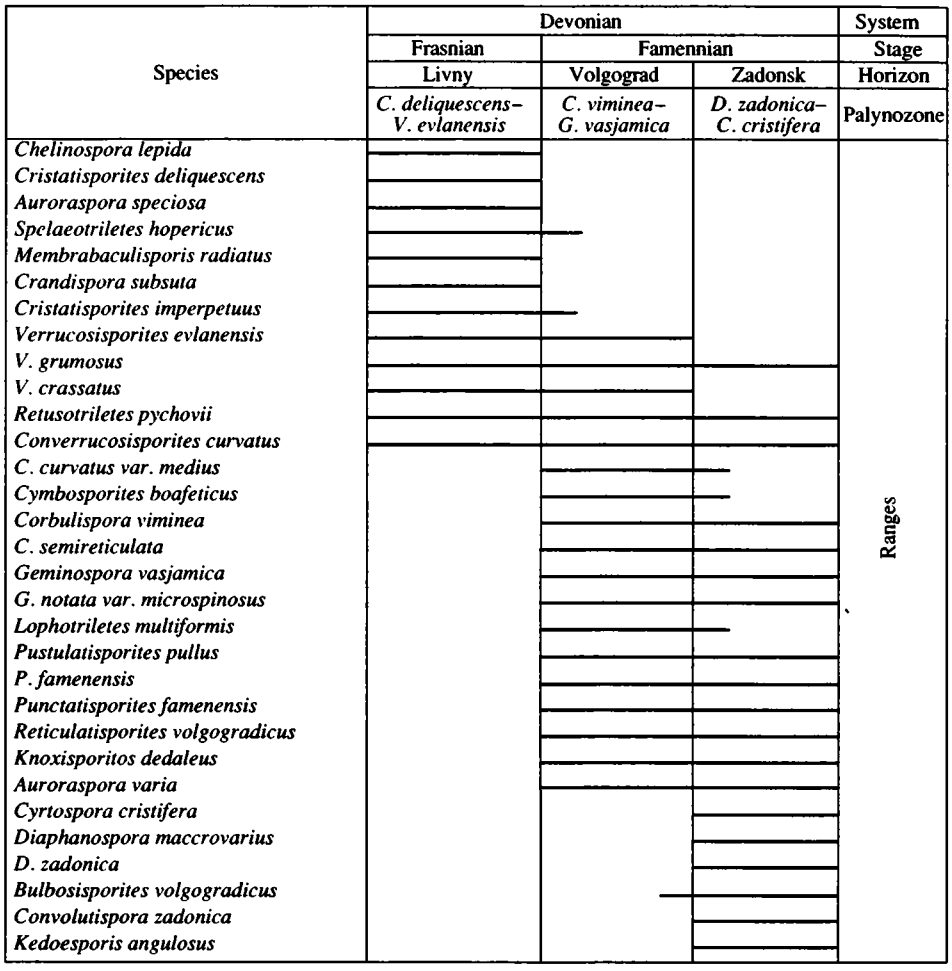

Fig. 7. Stratigraphic ranges of characteristic spores from the Frasnian–Famennian boundary interval of the Volga region.

In the Volga River region, the upper boundary of the Volgograd Horizon is less recognizable than the lower one by means of palynological data (Fig. 7). The majority of spore taxa transit into the Zadonsk Horizon (the *Diaphanospora zadonica–Cyrtospora cristifera* Palynozone) that is composed of marls, mudstones, and limestones. Characteristic species of the Zadonsk Horizon are *Diaphanospora zadonica* (Naum.) Avkhim., *Cyrtospora cristifera* (Luber) Van Der Zwan, *Bulbosisporites volgogradicus* (Naz.) Obukh., *Convolutispora zadonica* (Nekr.) Obukh. et Nekr., *Kedoesporis angulosus* (Naum.) Obukh., and some others.

As shown above, most essential changes in spore assemblages take place at the base of the Volgograd Horizon (the Linevo–Umet sequence). This level was chosen for the boundary between the Frasnian and Famennian stages (Mantsurova, 1987).

Distribution of spore taxa in the Volgograd beds recovered by Borehole 44-Krasnyi Yar and Borehole 3-Lomovo is shown in Tables 1 and 2. Sampling levels are specified in the section description.

CONODONTS

V.G. Khalymbadzha studied conodonts from the middle part of the Volgograd Horizon penetrated by Borehole 11 in the Lomovo area (interval 2612–2630 m, Fig. 4). The conodont assemblage consists of *Palmatolepis triangularis* Sannem., *P. subperlobata* Sannem., *P. delicatula clarcki* Ziegl., *Icriodus alternatus* Brans. et Mehl., *I. cornutus* Sannem., *I.* sp., and *Polygnathus* sp. (Table 4). A similar assemblage was macerated from Borehole 13-Lomovo (interval 2623–2632 m, Sample 3). It includes *Palmatolepis triangularis* Sannem., *P. subperlobata* Sannem., *P. delicatula clarcki* Ziegl., *Polygnathus* sp., and *Icriodus* aff. *cornutus* Sannem. The Volgograd assemblage of conodonts is characteristic of the middle subzone of the *P. triangularis* Zone owing to the presence of *Palmatolepis delicatula clarcki* that appears at the base of the subzone. All other species of the assemblage are typical of this subzone. *Palmatolepis delicatula delicatula* appears in its lower part and above. *Icriodus cornutus* occurs in the middle part of the subzone and higher. The assemblage has no species, which have their first occurrence level in the upper sub-

Table 1. Distribution of spore taxa in the section of Borehole 44 of the Krasnyi Yar area

Horizon, Depth interval, m, Sample number, Spores, % / Species	Volgograd							
	2296–2300		2304–2308	2313–2317	2323–2326	2329–2333	2338–2340	2400–2405
	1	1a	2	6	10	11	14	23
Stenozonotriletes definitus	14	11	9	14		5	9	
Converrucosisporites curvatus	12	9	3	9	9	5	8	33
Pustulatisporites pullus	9	8	11	15	30	8	15	12
Punctatisporites famenensis	3	5	4	2	4	5		
P. solidus	12							
P. minor	7					3		
Geminospora vasjamica	5	7	6	7	4	6	7	12
G.notata var. *microspinosus*	7	8	9	11	7	10	8	10
Retusotriletes pychovii	5	7	8	10		9	5	
Auroraspora varia	3		2		3	2	2	6
Corbulispora viminea	1		1	4			2	1
Lophotriletes multiformis		3	3		6	3	4	6
L. rugosus		9				5	3	
Cymbosporites boafeticus		7	2		2		3	
Leiotriletes trivialis	5	6	10	10	10	9	9	4
Archaeozonotriletes variabilis		3	3	2		3	2	
Hymenozonotriletes pallidus			6			3	9	
Diaphanospora rugosa					1	2		2

zone. The conodont *Palmatolepis triangularis* Zone is referred to the Famennian Stage according to the resolution of the International Subcommission on Devonian Stratigraphy.

BRACHIOPODS

Brachiopod distribution in the Volgograd Horizon was studied by V.I. Shevchenko (Karpov *et al.*, 1962; Dan'shina *et al.*, 1995). The lower part of the horizon contain species unknown from the underlying deposits, such as *Cyrtospirifer lomovensis* Schev. msc., *C. umetovensis* Schev. msc., *Athyris lomovensis* Schev. msc., *Lingula kotovensis* Schev. msc., *Liorhynchus kotovensis* Schev. msc., and *Adolfia* sp. nov., as well as rare poorly preserved *Theodossia* sp. Higher in the section there, appear *Cyrtospirifer asiaticus* Brice and *C.* ex gr. *asiaticus*, which become widespread in the overlying Zadonsk Horizon.

Among other fossils from the Volgograd deposits, there are known ostracodes, foraminifers, stromatoporoids, scolecodonts, acritarchs, and algae. Remains of ostracodes, foraminifers, and stromatoporoids occur mostly in the reworked clasts of the Livny deposits. Calcareous algae are represented by *Girvanella ducii* Weth. and *Rothpletzella devonica* (Masl.) of a wide stratigraphic range, or by accumulation of charophytes *Umbella bella* Masl. (Dan'shina *et al.*, 1995).

In the Volga River region, the Volgograd Horizon was reliably established in the Umet–Linevo paleodepression and in the Antipovo–Shcherbakovo zone of ancient uplift (Chibrikova and Nazarenko, 1974; Mantsurova, 1987, 2000) (Fig. 6). Analogs of the Volgograd Horizon may be recognized in the depth interval accessible for drilling along margins of the North Caspian depression. In the Volga–Ural region, the Volgograd Horizon was reliably established in the Volgograd and Saratov areas, Southern Urals, and the Cis-Urals region.

Figures 8 and 9 show the distribution areas of the Volgograd Horizons and its age analogs (the *Corbulispora viminea–Geminospora vasjamica* Palynozone) in the Russian platform and the Urals (Chibrikova and Nazarenko, 1974; Nazarenko *et al.*, 1993; Avkhimovich *et al.*, 1993; Obukhovskaya, 2000; and other). The correlatives of the Volgograd Horizon are the Domanovichi Horizon and the lower part of the Kuz'michevo Beds in the Pripyat' depression, the Pakul Beds and the lower

Table 2. Distribution of spore taxa in the section of Borehole 3 of the Lomovo area

Horizon, / Depth interval, m, / Sample number, / Spores, % / Species	Zadonsk		Volgograd					
	2595–2603	2603–2606	2606–2611		2611–2618	2618–2623	2623–2629	
	1	5	9	12	16	21	24	27
Converrucosisporites curvatus	12	6	17	9	13	20	7	8
Pustulatisporites pullus	11	10	11	12	10	14	13	7
Punctatisporites minor	7	1		4	2	3		
P. famenensis	5	2	3	4	2	1		2
P. solidus	4	5	2	2		1	4	
P. nigratus	1			1		0.5	2	1
Reticulatisporites perlotus	7	3		5		3	8	5
R. volgogradicus	2	5		4	2	1		1
R. retiformis			2		1			
Retusotriletes communis	7		4	10	2	10	4	
R. pychovii		5	5		2	1	7	7
Diaphanospora rugosa	7	8		0.5				2
Archaeozonotriletes variabilis	6	6	2	5	2	7	3	6
A. variabilis var. *insignis*					8			2
Stenozonotriletes definitus	6	8	7	2	4	8	10	6
S. laevigatus			3	2	1			1
S. pumilus		2			1			
Hystricosporites hamulus	4	4		1		1		
Kedoesporis livnensis	3	3	2	6		0.5		1
Bulbosisporites volgogradicus	2							
Grandispora famenensis	2	1	1	1				
Convolutispora zadonica	2							
Geminospora vasjamica	1		2	2			1	1
G. notata var. *microspinosus*	4	3	5	8	5	2	2	7
Lophozonotriletes aff. *torosus*	1			1	7			
Knoxisporites dedaleus		4				5		2
Verrucosisporites grumosus		3	2		1		3	
V. evlanensis			2			2		1
V. crassatus			2			3	2	1
ëorbulispora semireticulata					1	1	1	
Spelaeotriletes hopericus			0.5	1		1		3
Lophotriletes multiformis						1		
Dictyotriletes annae			2	1			1	
Cymbosporites boafeticus			2		1		2	4
Auroraspora varia			5	2	6			10
Lophotriletes multiformis			3		1			
Archaeozonotetraletes alius	1		2	1	2			1
Acritarcha				4			4	8

Table 3. Percentages of spores taxa most frequent in the Volgograd Horizon

Species	Quantity, %	
	maximal	average
Pustulatisporites pullus	40.0	12.0
Converrucosisporites curvatus	33.0	8.0
Stenozonotriletes definitus	25.0	7.5
Punctatisporites minor	20.0	3.5
P. famenensis	18.0	2.9
P. solidus	15.0	5.0
P. nigratus	4.0	0.5
Hystricosporites hamulus	18.0	2.1
Auroraspora varia	12.5	4.5
Geminospora vasjamica	12.0	6.0
G. notata var. *microspinosus*	15.0	6.5
Retusotriletes pychovii	12.3	6.4
R. communis	7.0	1.4
Leiotriletes trivialis	10.0	4.5
Diaphanospora rugosa	8.0	1.2
Archaeozonotriletes variabilis	8.4	4.0
Reticulatisporites perlotus	7.9	2.4
R. volgogradicus	9.1	3.6
Cymbosporites boafeticus	7.6	2.5
Lophotriletes multiformis	7.0	0.9
Corbulispora viminea	7.0	1.7
Verrucosisporites crassatus	5.5	0.9
V. grumosus	3.0	0.9
V. evlanensis	2.7	0.7
Acritarcha	90.0	8.0

Table 4. Distribution of conodonts in the section of Borehole 11 of the Lomovo area

Horizon, Depth interval, m, Sample number / Species	Volgograd								
	2612–2622					2622–2630			
	29	29a	30	34	35	40	48	49	49a
	+	+							+
P. delicatula delicatula Sannem.	+		+						
P. delicatula clarcki Ziegl.	+				+				+
P. subperlobata Sannem.	+	+							+
Polygnathus foliatus Bryant.	+								
P. ex gr. *brevilaminus*		+		+	+	+	+	+	+
Icriodus cornutus Sannem.							+		
I. alternatus Brans. & Mehl.									+
I. aff. *alternatus*							+		

Stage	Conodont zone	Spore zone	Horizon	Volga region: Horizon	Volga region: Unit (Formation)	Dnieper-Donets depression: Horizon	Dnieper-Donets depression: Beds	Pripyat depression: Horizon, beds	Timan-Pechora province: Horizon	Timan-Pechora province: Beds (Formation, Unit)	The Urals: (Formation, Unit)
Famennian	*P. crepida*	*D. zadonica–C. cristifera*	Zadonsk	Zadonsk		Zadonsk	Oleshnya	Zadonsk	Zadonsk		Third unit / Kiya Formation (lower part)
Famennian	*P. triangularis*	*C. viminea–G. vasjamica*	Volgograd	Volgograd	Umet		Leskovo	Kuz'michevo	"Under-Zadonsk"	Izhem Formation	Zilair Formation (lower part) / Second (Vazyam) Unit
					Linevo		Pakul	Domanovichi		Savinov Bor Unit	?
Frasnian	*P. linquiformis*	*C. deliquescens–V. evlanensis* subzone *G. subsuta*	Livny	Livny		Livny		Livny	Livny	Ukhta Formation	First unit / Askyn Formation

Fig. 8. Correlation of the Frasnian–Famennian boundary deposits in the Russian platform and the Urals based on palynological data (Nazarenko *et al.*, 1993).

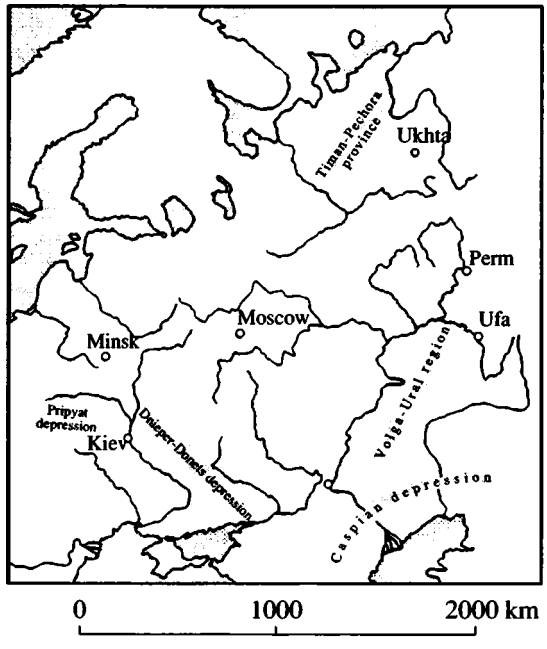

Fig. 9. Distribution areas of the Volgograd Horizon in the Russian platform.

part of the Leskovo Beds in the Dnieper–Donets depression, the lower Famennian beds below the Zadonsk Horizon in the Timan–Pechora region, and the Vazyam Sequence of the Zilair Formation in the Urals.

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Serpukhovian Stage Stratotype in the Zabor'e Quarry, Part 1: Lithofacies Characterization

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Received February 7, 2001; in final form, May 15, 2001

Abstract—The lithofacies characterization of beds in the Serpukhovian Stage stratotype, the Zabor'e Quarry site near Serpukhov, is presented. Ten members, boundaries of which do not coincide with those of local subdivisions formerly distinguished here, are established in the section. It is shown once again that the topmost portion of Venev Horizon near the section base is karstified. Above the karstified rocks, several levels of sedimentation basin shoaling, which are marked by rhizomorphs and indications of fresh-water diagenesis, are recognized. The first level crowns the lower third of the Tarusa Horizon, the second one is in the upper part of the latter and considered to be its upper boundary, and the third level in the upper third of the lover Steshev Subformation may characterize two nearby events of shoaling. Parallel to aridity indications enhancing upward in the Zabor'e and Novogurovskii sections, rocks of that subtidal reveal storm-born marks in subtidal facies. As is established for the first time for the Okaian–Serpukhovian interval, the high-energy grainstones and intertidal laminites with storm-born marks appear in the terminal part of the lower Steshev sequence. Thus, the transition from the humid climate of the Okaian–Tarusa time to the more arid climate of the Steshev time could be accompanied by enhanced movements of air masses that triggered development of storm sedimentation. The last conclusion needs however to be verified in other sections of the study region.

Key words: Serpukhovian Stage, stratotype, lithofacies, rhizomorphs, limestones, marls, mudstones, sedimentation, diagenesis.

INTRODUCTION

This study appears to be actual because of necessity to select and investigate in detail a concrete stratotype for the Serpukhovian Stage, the candidate subdivision for the global chronostratigraphic scale (*Stratigraficheskii...*, 1992. Addendum, paragraph 1.1). An appropriate section is that of the Zabor'e Quarry, where the stage is exposed throughout its section preserved in the stratotype area. The Zabor'e Quarry exists as long as 150 years, being known all the time as a main source of fossils of the Serpukhovian age.

Nikitin (1890) distinguished the Serpukhovian Stage in the range of *Spirifer trigonalis* beds in the stratigraphic scheme of A.O. Struve and pointed out that the Zabor'e section is one of other most typical sections in the vicinity of Serpukhov (Podmokloe, Luzhki, and Sknizhka River), which are well characterized in the paleontological aspect.

Shvetsov (1932) did not mention the Zabor'e Quarry among type localities of the Serpukhovian Stage, because its rock sequence was poorly exposed, as I think, in the 1920s. He published however the geological column of that locality, its description, and even photographs of the section in his later works (Shvetsov, 1940, 1948) presenting results of field investigations in type areas of the stage in 1939. One more column characterizing the Zabor'e section was published by Rauser-Chernousova (1948, p. 29). The section base was likely situated at that time in a lower part of the Tarusa Horizon (in Bed 3 or 4), and the top was at the level of beds 26–28 of the currently exposed succession (Fig. 2).

Despite the fact that rock sequence observable now in the Zabor'e Quarry is twice thicker than that of 1948, the structure of its upper portion is inadequately known at present. Osipova and Bel'skaya (1965a, 1967, 1975; Osipova *et al.*, 1972) who investigated lithology, facies, and fossil content of Serpukhovian Stage in the vicinity of Serpukhov, the Zabor'e section included, unfortunately published its characterization without bed-by-bed description of beds and columns of particular exposures. Schematic columns of that section have been published in some later works (Barskov and Alekseev, 1979; Makhlina *et al.*, 1993; Skompsky *et al.*, 1995), but its bed-by-bed description can be found only in the Guide to Field Excursion of 1998 that is unpublished and inaccessible for a wide auditorium.

Rock samples for this study have been collected during field seasons of 1999–2000 and used to prepare 55 thin sections of vertical orientation (sampling levels are marked by arrows in Fig. 2). Additional 350 thin sections from many limestone and marl beds of the quarry were kindly donated for investigation by N.B. Gibshman. Microtextures of the rocks are described below in terminology of Danham, Embry and

Occurrence frequency of grains counted in thin sections with average area of 9 cm^2

Occurrence frequency	Grainstone and packstone	Wackestone	Mudstone
Grains of 1–2 mm in size			
Single	1 grain per a thin section	1 grain per a thin section	1 grain per a thin section
Rare	up to 2 grains/cm^2	2–5 grains per a thin section	2 grains per a thin section
Frequent	2–7 grains/cm^2	up to 4 grains/cm^2	up to 1 grains/cm^2
Abundant	>7 grains/cm^2	>4 grains/cm^2	>1 grains/cm^2
Grains of 0.2–1 mm in size			
Single	1–3 grains per a thin section	1 grain per a thin section	1 grain per a thin section
Rare	up to 3 grains/cm^2	up to 1 grains/cm^2	2 grains per a thin section
Frequent	3–10 grains/cm^2	1–6 grains/cm^2	up to 2 grains/cm^2
Abundant	>10 grains/cm^2	>6 grains/cm^2	>2 grains/cm^2
Grains of 0.05–0.2 mm in size			
Single	1 grains/cm^2	1 grains/cm^2	1–3 grains per a thin section
Rare	2–7 grains/cm^2	2–5 grains/cm^2	up to 2 grains/cm^2
Frequent	8–20 grains/cm^2	6–15 grains/cm^2	2–7 grains/cm^2
Abundant	>20 grains/cm^2	>15 grains/cm^2	>7 grains/cm^2[22]

Klovan, and terms of Choquette and Pray (Tucker and Wright, 1990) are used to characterize the types of pores. Horizontal jointing or foliation produced by compaction of sedimentary rocks is termed here as fissility (*Tolkovyi...*, 1977) that corresponds to primary foliation in understanding of Russian geologists. The term neomorphism is applied to those secondary crystalline textures, the complex origin of which could include both the recrystallization and polymorphic transformation of substance, e.g., the aragonite–calcite transition (see Kabanov, in press, for more details). Other terms are explained in the text when necessary. Skeletal grains with an average size of 0.05– 2 mm have been counted in thin sections with due account for "precipitate dissolution by mud fraction" in accord with the presented Table.

ZABOR'E SECTION

The Zabor'e Quarry is situated 2 km away from the left bank of the Oka River in southeastern outskirts of Serpukhov near the Mirnyi Settlements (54°54′ N, 37°27′ E, Fig. 1). The quarry is out of mining now. The used numeration of beds from the base upward is that of Barskov and Goreva (see in Barskov and Alekseev, 1979; Skompsky *et al.*, 1995). According to results obtained, the section is divisible in ten members (Fig. 2). All beds that compose the members are described below.

First Bench

Visean Stage

Venev Horizon

Member 1 (beds 1, 2) includes gray fine-grained limestone (Bed 1) that has been exposed on the quarry bottom, an interval of talus (1 m), and lower 1.3 m of the currently exposed section (Bed 2). The latter is composed of karstified inhomogeneous mottled limestone with diverse dissolution caverns, rhizocretions of the *first root horizon*, and microscopic indications of fresh-water diagenesis. All the rocks represent the *Muratovo paleokarst profile* (Kabanov, in press). Microtextures correspond to those of packstone and wackestone-packstone of foraminifer-ostracod composition; rocks are fine-grained and bioturbated. In two thin sections, their coarse-grained areas reveal presence of *Calcifolium okense* Schwetz. (Plate I, Fig. 1; Kabanov, in press, Plate I, Figs. 2–4). The total thickness is about 2.5 m.

Serpukhovian Stage

Tarusa Horizon

Member 2 (beds 3, 4): light gray thick-bedded limestone with moderately abundant *Zoophycos* traces and with spaced *Stigmaria* appendices near the top. Under microscope, the rock corresponds to fine-grained bioclastic packstone and wackestone. Skeletal material is intensively crushed. The member consists of four plates of darker and harder, unclearly mottled limestones, which are separated by lighter and softer, slightly fissile limestone interlayers. Each bed consists of two such plates. Transition between soft and hard limestones is gradual everywhere; boundaries between plates are locally indistinct and can be almost invisible. The basal clay part of the member is highly compact and fissile, getting darker in places, where clayey material has been redeposited from the underlying weathering profile (Kabanov, in press). Beds 3 and 4 are separated by clayey foliated limestone. The member is 1.35 m thick.

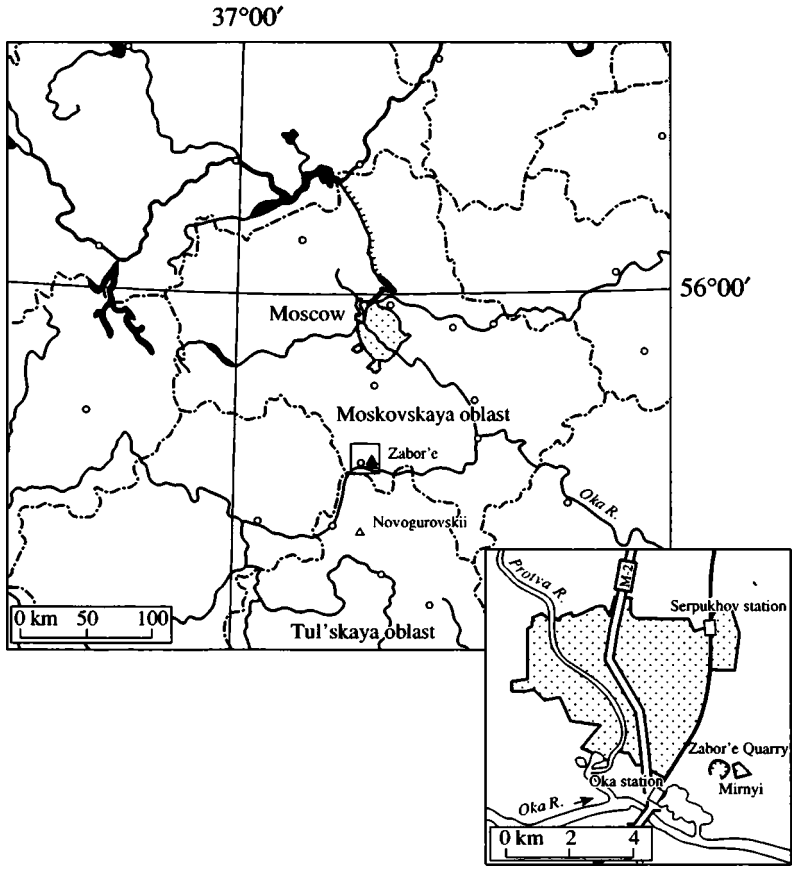
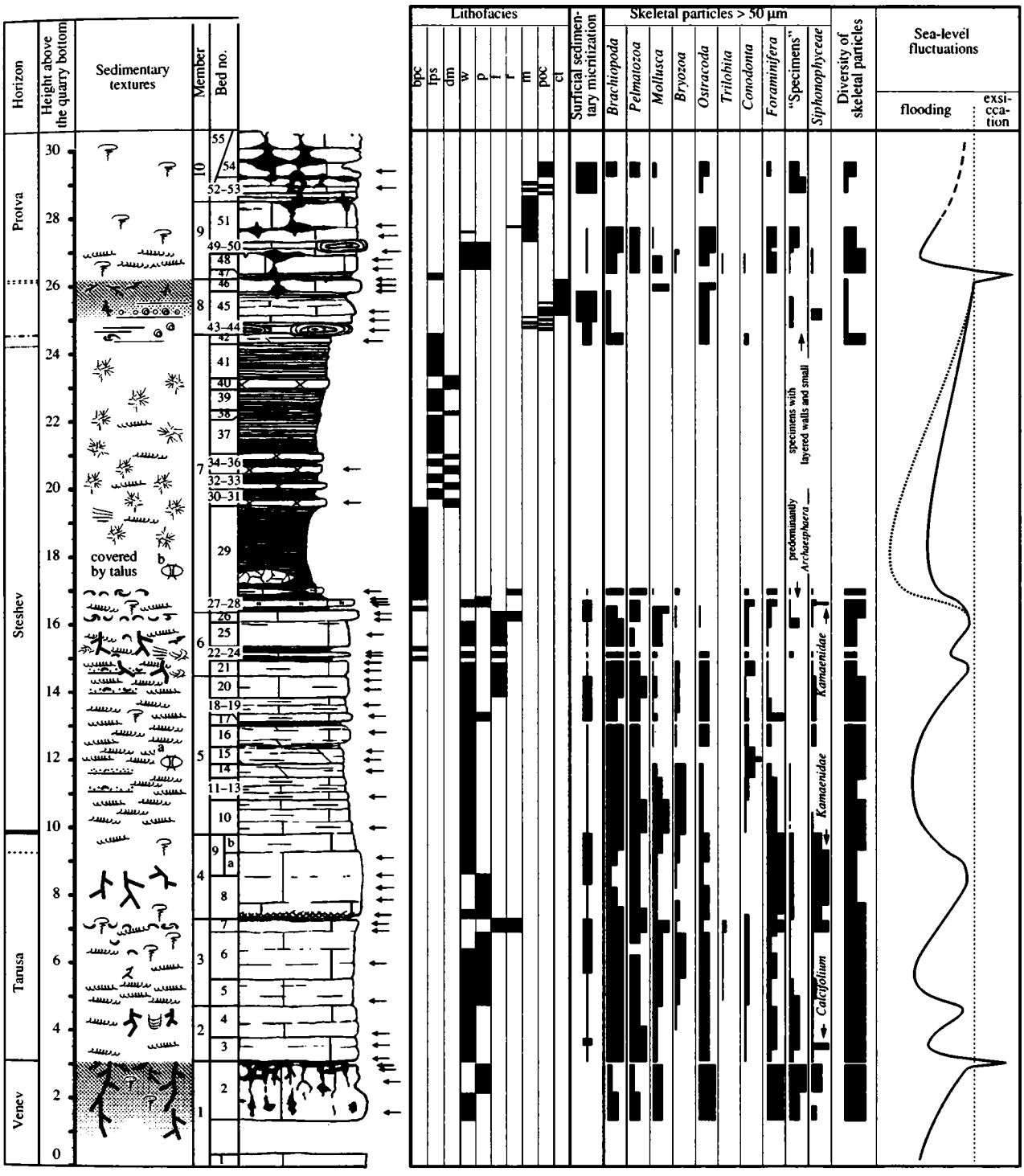


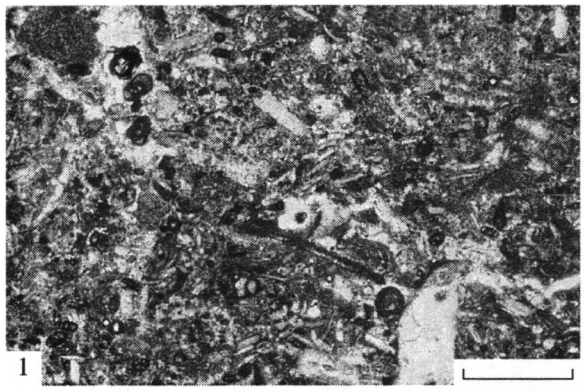
Fig. 1. Geographic localities of the Zabor'e and Novogurovskii sections.

Bed 3 (0.55 m): two plates of cream-gray limestone, which are nearly equal in thickness. Abundant biomolds (cavities after dissolved skeletal particles) and aggregates of large brachiopod fragments are characteristic. Under microscope, this is fine-grained packstone-wackestone with local areas of compaction microtextues. Skeletal material is intensively crushed except for foraminifers, which are mostly intact. A soft portion of the lower plate corresponds under microscope to the crinoid-foraminifer-brachiopod wackestone with secondary fluidal patterns of compaction origin (Plate I, Fig. 2). In a dark central spot of the upper plate, the rock corresponds to coarse-grained bioclastic *Calcifolium* wackestone-packstone (Plate I, Fig. 3) that is partially neomorphic, lacking signs of compaction. Cavities of the early generation (synsedimentary pores inside and between grains and biomolds after aragonitic shells) are filled in with equant sparite. Some echinoderm bioclasts are nearly black because of microscopic opaque inclusions.

Bed 4 (0.8 m): gray homogeneous bioturbated limestone similar to that of Bed 3. A rather distinct mottling

Fig. 2. Lithofacies characterization of the Zabor'e section: (1) limestone; (2) clayey limestone; (3) limy marl; (4) dolomitic marl; (5) plastic clay; (6) flaggy palygorskite shale; (7) cherty concretions; (8) levels of phosphatic nodules after Shvetsov (1948) (a) and after Alekseev (b); (9) rhizocretions and *Stigmaria* remains, large (a) and small (b); (10) coquina; (11) graded storm deposits with predominant (a) small (<2 mm) and (b) large (>2 mm) skeletal remains; (12) intertidal deposits of Member 8 with fine horizontal lamination (a) and with lenses of Siphonophycaceae-peloid-oncoid grainstone (b); (13) boundary between Tarusa and Steshev horizons (a) accepted formerly (Shvetsov, 1948; Makhlina *et al.*, 1993) and (b) suggested in this work; (14) boundary between Steshev and Protva horizons (a) after Makhlina *et al.*, (1993), (b) after Barskov and Alekseev (1979), and (c) suggested in this work; (15) profile of subaerial exposition; (16) pre-Moscovian karstified complex; (17) barite-limonite nodules; (18) *Zoophycos*, (19) *Diplocraterion*, (20) *Teichichnus*, (21) *Vermichnus*, (22) *Planolites*, and (23) *Fucoides* s.l. taxa of ichnofossils; (24) swirling bioturbation patterns; (25) development of sedimentary micritization, (a) intense (>60% of skeletal particles is micritized), (b) considerable (25–60% of skeletal particles is micritized), (c) weak (5–25% of skeletal particles is micritized), and minimal (<5% of skeletal particles is micritized); (26) occurrence frequency of small skeletal particles (a) abundant, (b) frequent, (c) rare, and (d) single as graded in the Table; (27) diversity of skeletal particles (a) very high, (b) high, (c) moderate, and (d) low; (bpc) black plastic clay; (fps) flaggy palygorskite shale; (dm) dolomitic marl; (m) mudstone, (w) wackestone, (p) packstone, (f) floatstone, (r) rudstone, (poc) peloid-oncoid calcarenite, and (ct) calcrete lithofacies of limestone or marl.


Horizon
Height above the quarry bottom
Sedimentary textures
Member
Bed no.
Lithofacies
bpc
fps
dm
w
p
f
r
m
poc
ct
Surficial sedimentary micritization
Skeletal particles > 50 μm
Brachiopoda
Pelmatozoa
Mollusca
Bryozoa
Ostracoda
Trilobita
Conodonta
Foraminifera
"Specimens"
Siphonophyceae
Diversity of skeletal particles
Sea-level fluctuations
flooding
exsiccation
Protva
Steshev
Tarusa
Venev
covered by talus
specimens with layered walls and small
predominantly Archaesphaera
Kamaenidae
Kamaenidae
Calcifolium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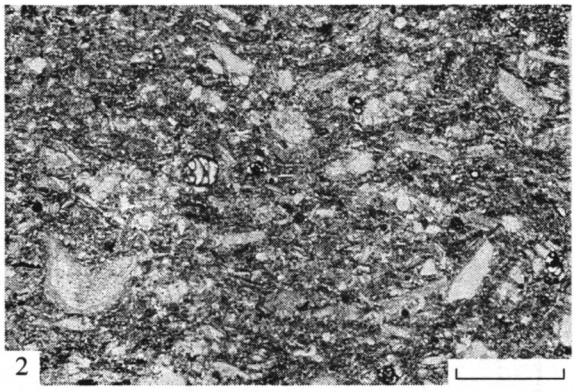
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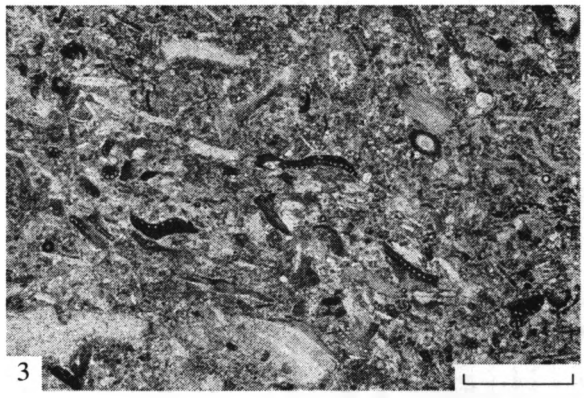
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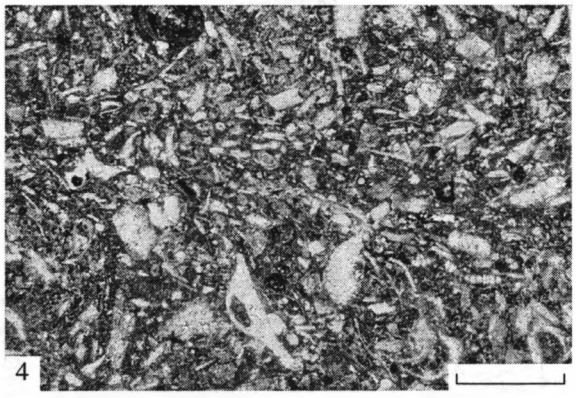
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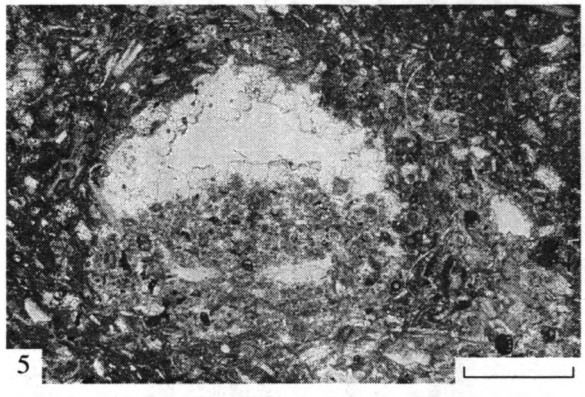
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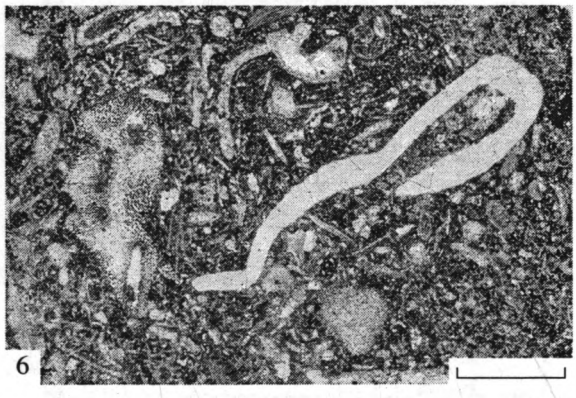
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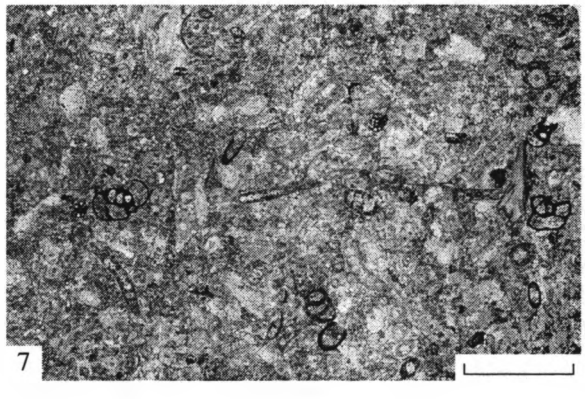
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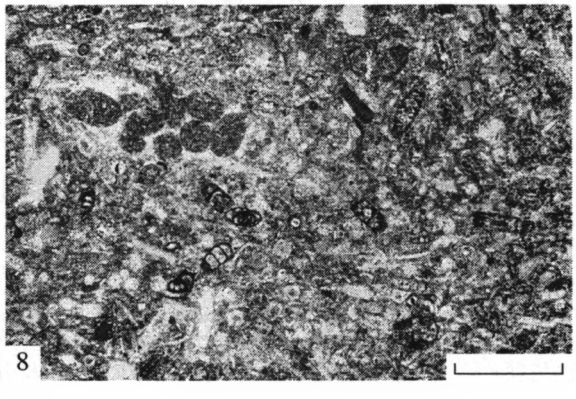
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after *Stigmaria* appendices and bioturbation traces are characteristic of the lower plate. Traces of *Zoophycos* and more rare *Diplocraterion* are distinguishable. Thin-shelled abundant brachiopods are buried, sometimes in the life position, throughout the bed. Biomolds after molluscan shells occur as well. Under microscope, the rock corresponds microbioclastic wackestone with a characteristic bioturbation texture of *Zoophycos* type. In general, skeletal fragments are crushed. Matrix is partially neomorphic. Pores inside skeletal fragments and biomolds are filled in, partially or completely, with equant sparite.

Bed 4 encompasses the *second root horizon* with *Stigmaria* appendices, the long subvertical tubes of about 10 mm in diameter, which are sometimes empty, with a halo of weak alterations in surrounding rock. The tubes are characteristic of the entire thickness of Bed 4. They are identical to appendices of some *Stigmaria* beds in the Oka Superhorizon of the southern limb of Moscow syneclise (Gekker, 1980), being distinct owing to a greater in-between spacing.

Thin sections from this member, like from many muddy lithofacies occurring higher in the section, reveal matrix-engulfed allochems in areas of dense micrite (Plate 1. Figs. 3, 6,). In these cases, degrading allochems show partial amalgamation and blurred boundaries with surrounding micrite. This feature can be likely attributed to internal micritization (May, Perkins, 1979). Similar transformations are widespread in muddy lithofacies of Middle Carboniferous limestones of the Moscow syneclise (Kabanov, 2000).

Member 3 (beds 5–7): light gray thick-bedded limestone indistinctly banded downward because of intense development of *Zoophycos* structures, which decrease in abundance upward in contrast to large skeletal fragments whose abundance grow in this direction. The uppermost Bed 7 or "lower coquina" bounded by hummocky surface at the top is most rich in intact macrofossil remains and their fragments. Under microscope, the rock mostly corresponds to bioclastic wackestone with fine to medium grain size clasts of skeletal material. Abundance of large bioclasts increases upward parallel to decreasing proportion of microbioclastic micrite matrix (primary muddy fraction of sediment). At the base of bed 7, microtexture grades into that of floatstone-rudstone. The member is 2.1 m thick.

Bed 5 (0.6 m): hard limestone more porous and soft near the base, where dark *Zoophycos* bands are abundant. Dark mottles with biomolds after molluscan shells are confined to the upper bed part. Under microscope, the rock represents polybioclastic wackestone-packstone locally enriched in coarse skeletal fragments, which are greater than 2 mm across (Plate I, Fig. 4). Dense wackestone exhibits the matrix-engulfed bioclasts. Some, slightly micritized bioclasts reveal microborings filled in with micrite and discontinuous destructive (developing inside from the surface) rims of micrite.

Bed 6 (1.3 m): gray, comparatively hard limestone with local darker and blurred, horizontally extended mottles. The rock is softer and pelitomorphic near the base. Swirling bioturbation patterns of irregular arrangement are interrupted by *Zoophycos* mottles. Transition to Bed 7 is gradual with stylolites. Under microscope, the inhomogeneous polybioclastic to micrite texture consists of dense wackestone areas and accumulations of skeletal material corresponding to inequigranular or coarse-grained packstone. Some bioclasts are rounded, insignificantly micritized, displaying microborings. Within dark mottles, the packstone matrix is less dense, consisting of shreds or clots set in equant sparite. The latter partially fills in cavities as cement or forms neomorphic areas replacing the micritic matrix. In the softer rock, areas of the same primary microtexture with a lesser proportion of matrix are almost lacking sparite. The distinguished single tubes with peloid-grumous geopetal filling differ from surrounding mass in their color and texture (Plate I, Fig. 8). It is likely that these tubes are former openings left by roots. Since undisputable microscopic traces of roots have not been established between the aforementioned "root horizons," I believe that the tubes in question penetrated deep in sediment from the third root horizon.

Bed 7 (lower coquina; 0.2 m): light gray limestone with yellow mottles and abundant diverse macrofossils, which form the framework of rock in places of their clustering. Biomolds after diverse molluscan shells, brachiopods (representing mostly the genera *Productus, Antiquatonia, Eomarginifera*, and *Composita*), and bryozoans are abundant. Intact shells are rather frequent, representing about a half of skeletal remains. Many brachiopods occur *in situ* or are insignificantly

Plate I. Limestone microfacies of the Tarusa Horizon:

(1) Bed 2, 1.3 m below the top, brachiopod-ostracod-foraminiferal packstone-wackestone with a section of *Calcifolium okense* Schwetz. (2) The Tarusa Horizon base, bioclastic wackestone, prevailing horizontal orientation of grains and fluidal patterns represent a result of nonlithified sediment compaction. (3) A dark spot in the upper plate of Bed 3, inhomogeneous neomorphic bioclastic wackestone-packstone with abundant *Calcifolium* remains and cavities filled in by equant sparite. (4) The base of Bed 5, bioclastic packstone with abundant biomolds after molluscan shells, which are filled in by equant sparite; bioclasts are oriented along falcate filling-ins of *Zoophycos* spraythen. (5) The middle of Bed 6, section of a tube with geopetal sediment in polybioclastic wackestone. (6) An upper portion of Bed 7, bioclastic floatstone-rudstone; center to the left, a section of trilobite test; at the right, a micritized echinoderm fragment; matrix-engulfed bioclasts are visible. (7) An upper portion of Bed 8, neomorphic foraminiferal-kamaenid packstone. (8) The middle of Bed 9, foraminiferal-bioclastic wackestone with still abundant kamaenids and visible invertebrate boring filled in with fecal pellets (scale bars 1 mm in all figures).

displaced. Macrofossil remains are cemented by grainy pelitomorphic limestone with sporadic *Zoophycos* traces. Under microscope, the rock represents the brachiopod-molluscan floatstone-rudstone with abundant Siphonophycaceae thalli and sporadic trilobite remains (Plate 1. Fig. 6). Matrix is sometimes loose and neomorphic, but dense and engulfing bioclasts in other places. Nearly all biomolds are filled in with equant sparite. Many skeletal remains (biomolds included) are moderately micritized. As is visible under microscope, skeletal fragments are chaotically dispersed. The upper surface of Bed 7 is hummocky, ferruginate and somewhat brightened; A.V. Mazaev found here a shield-shaped *Syringopora* colony with corallites parallel to each other, attached to that surface, and implying its syngenetic lithification. Shreds of rusty foliated marl are locally preserved on the surface, but in general, this level experienced the late diagenetic alteration. The contact between beds 7 and 8 is locally marked by black smeary clay, ferruginate concretions (probably decomposed pyrite), and secondary cavities half-filled with brown and loose ferruginate lumpy clay that presumably is a weathering product of foliated marl. Because of alteration and unclear indications of subaerial diagenesis, it is hard to say whether that hummocky surface was a hardground or the underlying rocks represented a carr.

The second and third members look similar. The microscopic examination shows that main skeletal components, except for *Calcifolium okense* characteristic of Member 2 only, are inherited in any bed from the previous one. The diverse, distinctly autochthonous assemblages of macro- and microfossils from this interval of sediments lacking signs of hydrodynamic sorting suggest, together with the intense reworking of sedimentary material by detritovores, the normal marine low-energy sedimentation environments in a subtidal zone. The *Stigmaria* appendices, which appear in Bed 4, point to a short-term shoaling episode, during which a shallow sea floor was populated by rare lepidophytes. During the accumulation time of Bed 3, the ground was muddy, replaced afterward first by muddy-sandy (beds 4, 5) and then by muddy-shelly substratum (beds 6, 7). Such feature as micrite envelopes around bioclasts with microborings and higher abundance of Siphonophycaceae (Kamaenidae) remains in Bed 7 are clear indications of shoaling, the further development of which is recorded in the next Member 4. It is also possible that a brief subaerial exposition took place immediately after accumulation of Bed 7.

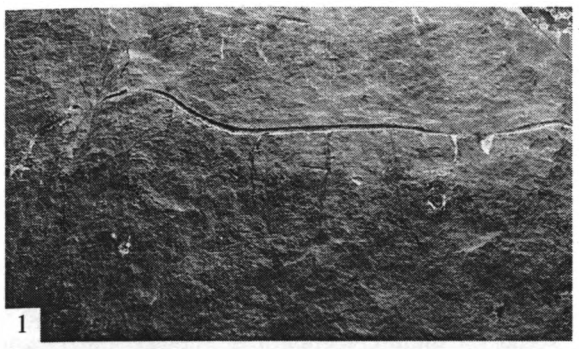
Member 4 (beds 8, 9): Gray fine-grained limestones of the member have blurred dark mottles and are comparatively hard, lacking *Zoophycos* remains, and nearly aphanitic in the upper part. In their microtextures, the rocks predominantly correspond to fine-grained packstone with abundant Siphonophycaceae remains (Kamaenidae, determination by N.B. Gibshman) and turn into wackestone near the top. Rhizocretions characteristic of the entire thickness of Member 4, except for the upper plate of Bed 9, represent markers of the *third root horizon.* Kamaenidae, the tubular septate structures in crystals showing a synchronous extinction, were attributed by Shvetsov (1948, p. 130) to Nodosinellae, which do not occur, in his opinion, above the Tarusa Horizon, but Gibshman found them in abundance in Bed 28 marking the middle of the Steshev Horizon. The member is 2.3 m thick.

Bed 8 (1.2 m): comparatively hard limestone with small vugs (0.5–1 mm) and indistinct horizontal mottles, which locally exhibit a sedimentary lamination that is outlined by subhorizontal orientation of allochems. Owing to absence of *Zoophycos* patterns, the rock looks homogeneous, nearly massive. Under microscope, it reveals the upward transition from basal polybioclastic packstone-wackestone of variable grain size to homogeneous fine-grained kamaenid packstone (Plate I, Fig. 7). Under microscope, skeletal fragments are of a low diversity. The perfect preservation of foraminifers, ostracods, and Kamaenidae implies their autochthonous origin. Other bioclasts are small, intensively reworked, and apparently allochthonous. The rock is irregularly neomorphic, lacking signs of compaction. The boundary between beds 8 and 9 is conventionally placed at the concentration level of rhizocretions in the third root horizon (see below).

Bed 9 (1.1 m): massive limestone lighter in color than that of Bed 8 and locally displaying either a sedimentary lamination or bioturbation; near the top, proportion of clay admixture is somewhat higher. A lower half of the bed (layer 9a) is amalgamated with Bed 8 to represent a single thick plate. Its upper portion 0.5 m thick (layer 9b) corresponds to another plate that again shows appearance of rare *Zoophycos* forms. Under microscope, this is foraminiferal-bioclastic wackestone lacking compaction signs and enriched in Kamaenidae and small biomolds after molluscan shells. Bioclasts

Plate II. Third root horizon and microfacies of the Steshev clayey limestones:

(1) Beds 8 and 9, a rhizomorph consisting of main stem (2 cm in diameter and about 1.3 m long) and perpendicular appendices branching downward. (2) The middle of Bed 14, bryozoan-brachiopod wackestone with a *Zoophycos* structure. (3) Bed 21, bioturbated aggregate of large skeletal fragments in the crinoid-brachiopod packstone-wackestone; like in Fig 4, black impregnation may represent a result of pyritization. (4) A dark mottle in the middle of Bed 25, brachiopod floatstone with intersected "chervotochinas" (warm tubes) and biomolds, around which the fine-grained half-decomposed pyrite is developed. (5) Bed 26 (upper coquina), neomorphic brachiopod rudstone with coarse-grained equant sparite cement. (6, 7) Two microfacies of Bed 28: (6) dolomitized wackestone with neomorphic spines (sponge spicules?); (7) porous kamaenid packstone (scale bars 1 mm in figs. 2–7).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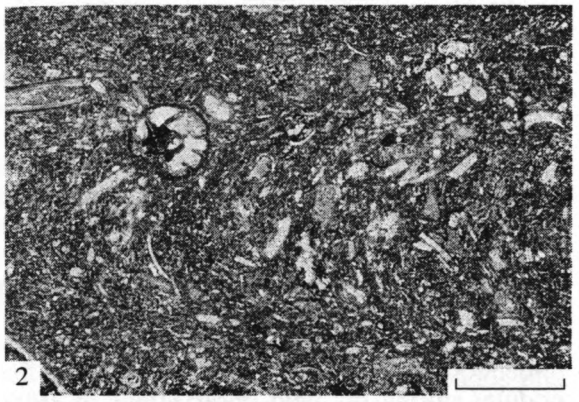
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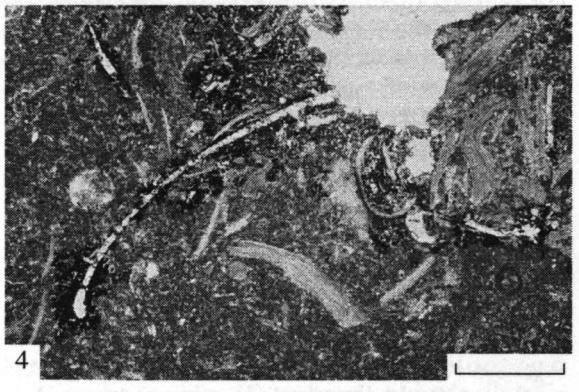
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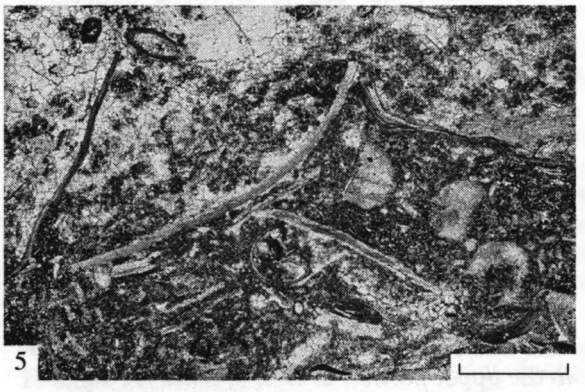
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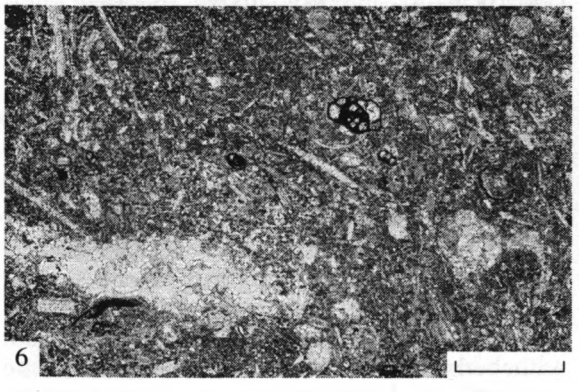
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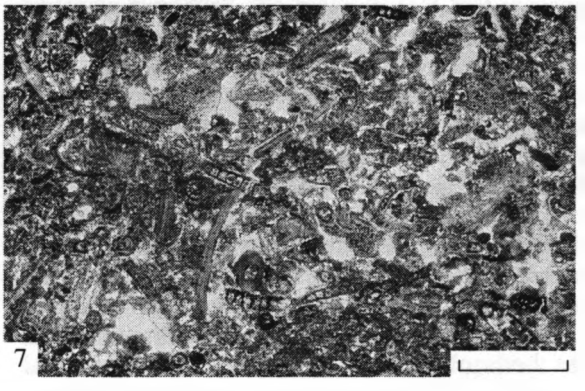
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engulfed by micritic matrix are quite widespread (Plate I, Fig. 8).

The third root horizon is marked by occurrence of rhizoconcretions and rhizocretions of variable orientation, which are up to 3–4 cm in diameter and locally preserve appendices branching under normal angle from the main stem (Plate II, Fig. 1). Around rhizomorphs, there are developed dark haloes and spots with microchannels and vugs, which have alveolate-tubercular walls. In addition, a vertical section in the lover part of Bed 9 exhibits a series of parallel imprints oriented under angle of about 45°, which resemble whips and represent appendices of *Stigmaria* of the second type after Gekker (1980). Judging from the root types, the corresponding vegetation consisted of two plant species at least, and that lycopods were comparatively rare, in distinction from similar root horizons of Okaian limestones, among other taxa, the taxonomic affinity of which remains to be established.

The impoverished composition of small skeletal remains, impaired bioturbation, mass abundance of uniform Kamaenidae and rhizocretions imply that Member 4 accumulated in an extremely shallow setting. Judging from neomorphic features of the rocks, which have dissolution fenestrae and vugs but are lacking signs of compaction (a criterion of early lithification), sediments experienced the early fresh-water diagenesis. Nevertheless, the absence of subaerial exposition surfaces means that the basin floor has never been exposed above the sea level for a noticeable period. The shoaling developed likely to the state of an intertidal zone or to appearance of a very shallow half-isolated lagoon or sea marsh. Plants that colonized this setting must characterize a peculiar ecotype resembling the present-day mangrove coast. I should mention here that idea to interpret *Stigmaria* from the Okaian fine-grained "rhizoid limestones" as remains of mangrove plants (Shvetsov, 1938; Osipova and Bel'skaya 1967, 1977; Osipova *et al.*, 1972) seems to be incorrect. According to my own observations, these limestones usually are components of the subaerial karstification profiles, and the plants most likely grew on land. Beginning from the upper portion of Bed 9, there are signs of sedimentation setting deepening: the appearance of *Zoophycos* and growing diversity of small skeletal remains of normal marine fauna.

Steshev Horizon

Following Shvetsov (1948), many researchers (Rauser-Chernousova, 1948; Barskov and Alekseev, 1979; Makhlina *et al.*, 1993) place the boundary between Tarusa and Steshev horizons in the Zabor'e section at the level of transition from monolithic blocky limestones of Member 4 to thick-bedded clayey limestones and marl of Member 5. Osipova and Bel'skaya (1965a) likely attributed to the Tarusa Horizon the basal limestone beds of the Steshev Horizon, which bear accumulations of *Dibunophyllum bipartitum* McCoy, but I cannot state this with confidence, because their work is lacking the column and description of the section under consideration. I propose to place the boundary between horizons at the level of maximum shoaling inside the Member 4. This would be compatible with the early idea of Shvetsov to choose boundaries between the Lower Carboniferous horizons at the maximum shoaling levels inside transitional intervals lacking adequate paleontological characteristics (Shvetsov, 1938, 1948; Osipova and Bel'skaya, 1965a). As the plate 9a is distinctly free of rhizomorphs and exhibits signs of incipient shoaling, I place the boundary in question at its base.

Member 5 (beds 10–20): The member is composed of fine-grained to aphanitic clayey limestone and gray marls, which have reddish to brownish tint. These bedded rocks are moderately fissile, having dark mottles and bands, displaying abundant structures *Zoophycos*, and enclosing interlayers with abundant detritus of fish skeletons. Under microscope, these are bioclastic mudstones, wackestones, floatstones, and rarer packstones. Small skeletal remains of normal marine fauna are diverse. Shvetsov (1948) referred in his work to a horizon of phosphorite nodules (from 1cm to 2 m in diameter) in a bed of platy and banded marl above the Tarusa Horizon, i.e., at the level of beds 14 or 15, as they are understood in this work. I failed however to find these nodules during my field observations of 1999–2000. The member is 5,2 m thick.

Bed 10 (about 1.0 m): gray homogeneous clayey limestone; the rock is bedded, bearing structures *Zoophycos*. In thin section from basal interval, the rock corresponds to floatstone-wackestone with molluscan, crinoid, and brachiopod clasts of variable size. Some skeletal particles are micritized at the surface. Usually biomolds are empty, containing sometimes an insignificant amount of equant sparite.

Second Bench

Beds 11–13 (0.8 m): rusty-gray limestone (reddish near the top); the rock is clayey, pelitomorphic, and fissile, displaying abundant *Zoophycos* traces. Lenticular and other accumulations of large shelly fragments (mostly of brachiopods) likely represent bioturbated storm concentrates. Under microscope, this is bioclastic mudstone with admixture of silty quartz grains and neomorphic spiny structures (sponge spicules?). Some bioclasts are silicified. Dispersed dolomite rhombs and crystal-molds (cavities after crystals, after dolomite in this case) are characteristic of this and higher beds.

Bed 14 (0.4 m): reddish gray fissile marl with *Zoophycos* and abundant fish remains; the rock is pelitomorphic, with laminae, lentils, and other segregations of bioclasts. Some bioclasts are silicified and visible by unaided eye as minute black inclusions. Boundaries with beds 13 and 15 are marked by thin laminae of compact clay. Under microscope, this is fine- to coarse-

grained bryozoan-brachiopod floatstone and wackestone (Plate II, Fig. 2) with dispersed cavities after dolomite crystals about 60 μm across. Nearly all bioclasts are angular. Indications of shallow sedimentation depth, such as algae remains or surface micritization are absent.

Bed 15 (0.55 m): mottled marl; the rock with abundant *Zoophycos* traces is reddish gray with dark gray mottles extended parallel to bedding. Many bioclasts are silicified, and the rock is compacted and fissile in general. In the basal interval, lenses of less compact gray limestone are bounded at the top by tubercular surface with postsedimentary extension cracks produced under lithostatic pressure or in response to syneresis. These lenses likely represent the firmground. Under microscope, the rock corresponds to polybioclastic wackestone with segregations of neomorphic bioclasts and/or biomolds with sparite filling, some of which originated after molluscan shells. Matrix is locally neomorphic, displaying grumous texture. The fossil assemblage is very diverse. Algae remains and events of micritization are sporadic. The destruction degree of bioclasts is higher than in beds 13 and 14.

Bed 16 (0.75 m): above the basal foliated lamina of calcareous clay, the rock corresponds to banded *Zoophycos* marl of brownish gray color. Limestone in the middle is darker and hard, almost lacking bands and slightly compacted. It grades downward and upward into marl resembling a kind of bed concretion. Many bioclasts are silicified. Cavities with alveolate walls seem to be formed after sedimentary structures (maybe after gypsum nodules), which moved aside sediment by growth. Under microscope, this is ostracod-crinoid-brachiopod wackestone with frequent conodonts.

Bed 17 (0.1–0.15 m): ocherous foliated calcareous clay with aggregates of bioclasts and *Zoophycos* forms.

Bed 18 (0.25 m): gray clayey hard limestone with swirling bioturbation patterns and *Zoophycos* forms. Near the top, thin fucoids (up to 5 mm in diameter) with alteration haloes are sometimes filled in with sparry calcite. Under microscope, this is fine-grained bioclastic packstone-wackestone with neomorphic segregations of spiny appearance (sponge spicules?) and biomolds after molluscan shells. Pores inside skeletal material and biomolds are filled in with equant sparite. The micritic to microbioclastic matrix is neomorphic in places of *Zoophycos* structures. Some bioclasts with endolithic microborings are slightly micritized. Between beds 18 and 19, there is a lamina of foliated clay (<0.05 m).

Bed 19 (0.55 m): gray clayey limestone that is moderately fissile, soft, and sometimes banded owing to ferruginate coloration of *Zoophycos* traces. The rock is enriched in detritus of fish remains. The bed is split in two plates 0.25 and 0.3 m thick. Under microscope, this fine-grained bioclastic wackestone reveals micritization of some echinoderm clasts and presence of scarce algae remains.

Bed 20 (0.75 m): banded *Zoophycos* limestone that is lighter in color and harder than that of Bed 19; the rock encloses lenses of coarse bioclastic material. In the middle of the bed, there are dark gray mottles. Brachiopod accumulations and solitary rugoses occur in the bed, the upper portion of which reveals presence of rare rhizocretions. Fenestrae and vugs of various scale, which could be related in origin to plant roots, are half-filled with sparry calcite. As is seen under microscope, the secondary pores are surrounded by haloes of fine-grained opaque matter (pyrite?). *Planolites* forms are confined to the top portion. Under microscope, the rock represents wackestone-floatstone composed of crinoid-brachiopod size-variable clasts and bearing abundant biomolds after thin-shelled bivalves and gastropods, and also rare Siphonophycaceae forms.

Member 5 is lacking clear indications of shallow-water sedimentation environments: micritization is occasional in its interval, rounded clasts are absent, and algae remains occur as single specimens. In contrast, *Zoophycos* forms are abundant. These features evidence in favor of normal marine sedimentation in a subtidal, comparatively deep zone. The member reveals appearance of a new sedimentary feature—the graded laminae disturbed by bioturbation. These are accumulations of large bioclasts and intact shells (coquinas), which are set in matrix, rest on a sharp erosion surface, and grade upward into a finer clastic material with a greater proportion of cementing mass (Plate II, Fig. 3). Structures of this kind originate by a storm roiling and redeposition, being especially characteristic of open-marine subtidal facies (Aigner, 1985). In the Novogurovskii section, the graded laminae appear approximately at the same level. In both sections, they have never been observed in the underlying Tarusa Horizon, while at the Novogurovskii locality, separate lenses of graded fine-grained material are recorded only in the Okaian beds. The fact that storm features are poorly developed or completely missed in places below the Steshev Horizon cannot be explained as a consequence of bioturbation, because they appear in *Zoophycos*-bearing strata displaying abundant bioturbation marks. Thus, provided that the storm features can be distinguished at the same level in other sections of the Moscow syneclise, we may suspect that the early Steshev time marked the developments of stormy weather under influence of strong winds in contrast to the wind-free climate of the Okaian–Tarusa period. Moreover, the storm features appear in the interval of both sections, where rocks bear abundant dolomite crystals or cavities left after them. It is also interesting that the Steshev time of progressive aridity growth (Osipova and Bel'skaya, 1965a, 1977; Makhlina *et al.*, 1993) coincides with the inferable period of stormy weather development in the region. Consequently, this local climatic change could be connected with the coeval global differentiation of climate and with development of Gondwana Glaciation.

Member 6 (beds 21–26): Limestones of the member are mostly hard, clayey, somewhat bituminous, with equant sparite cement. These unclearly mottled gray rocks with ocherous and reddish tints are intercalated with dark gray to black interlayers of foliated clay. Rhizocretions distinguishable in the member are similar to those of Member 4 and mark the *fourth root horizon*. It is likely that in this case we deal with two closely spaced root horizons. Large intact shells of gastropods and brachiopods are abundant in some beds. The latter represent the *E. lobate* assemblage. Bed 26 crowning the member corresponds to the "upper coquina" of a monotonous taxonomic composition that is mostly composed of brachiopod remains dominating over other fossils. Records of storm-influenced sedimentation are concentrated in the lower part and may be characteristic of the entire member. Ichnocoenosis is diverse (especially in dark clay beds), consisting of *Vermichnus, Planolites, Zoophycos*, and/or *Teichichnus* forms. Textures distinguishable under microscope, are those of wackestone, floatstone, and rudstone with a high proportion of intact shells. Micritization is a rare phenomenon. Small skeletal remains are rather uniform in composition. The member is 2.0 m thick.

Bed 21 (0.55 m): gray limestone; in the lower portion, the rock is yellowish gray, clayey and softer, displaying the compaction fissility and bearing *Zoophycos*, abundant brachiopods, solitary rugoses, and other macrofossils. In the upper half of the bed, limestone is hard, lacking compaction signs, and yielding brachiopod and gastropod shells. Small rhizocretions (up to 2 cm in diameter) are developed below the top. Biomolds after molluscan shells are filled in with sparite that forms sometimes the shell pseudomorphs. Rhizocretions with dark ferruginate haloes are lined inside with loose white or ocherous carbonate. Small (up to 5–8 cm) dissolution cavities are locally confined to rhizocretions. Under microscope, the rock corresponds to crinoid-brachiopod floatstone and wackestone with abundant conodonts and dispersed rhombohedric cavities after dolomite crystals about 50 μm across.

In the lower part of the bed, laminae of shelly and shelly-bioclastic concentrates (brachiopod rudstones) have sharp, sometimes erosional lower boundaries (Plate II, Fig. 6). These loose accumulations of poorly sorted shells and clasts have a low proportion of matrix, abundance of which gradually grows upward, and both features are characteristic of sediments deposited under influence of storm roiling and redeposition (Aigner, 1985).

Bed 22 (0.1 m): dark gray clay that is foliated, smeary, and soapy grades upward into marly shale. Transitions to beds 21 and 23 are gradual. The diverse ichnocoenosis includes *Zoophycos* and/or *Teichichnus, Planolites*, and *Vermichnus* forms. Abundant brachiopods and fish detritus are confined to bedding planes. Elongated convex accumulations of bioclasts likely represent coprolites of fishes.

Bed 23 (0.15–0.2 m): gray to dark gray clayey and bituminous hard limestone with diverse trace fossils (*Planolites, Zoophycos*, and/or *Teichichnus*), which are easily detectable owing to selective black coloration. The upper boundary is even and distinct. Under microscope, the rock represents ostracod and crinoid-brachiopod floatstone-wackestone with size-variable clasts and frequent conodonts. Admixture of fine-grained quartz has been detected near the top. The filling-in of biomolds consists of equant sparite. Matrix is locally neomorphic.

Bed 24 (0.1 m): dark gray to black soapy clay; transitions to beds 23 and 25 are gradual.

Bed 25: reddish gray limestone with *Zoophycos* forms and rhizocretions; organogenic traces are especially abundant in the middle of the bed. Rhizocretions have dark gray dens haloes, and some of them seem to be enlarged by dissolution. Characteristic of haloes are clusters of narrow (3–4 mm and less) undulating tubular structures ("chervotochinas"), which likely represent appendices of rhizocretions, being of the same generation with the latter. In mottles, there are well-preserved casts of gastropods. Near the base and top, the bed is colored lighter and slightly fissile in contrast to its massive non-compacted middle interval. Under microscope, this is brachiopod wackestone-floatstone with biomolds after gastropod shells. Large skeletal fragments are irregularly distributed, and many of them reveal the inner nidi of silicification. Fenestrae of the early generation, except for most tubular structures, are incompletely or completely filled in with equant sparite. Matrix is locally neomorphic. Opaque spots surrounded by ferruginate Liesegang rings likely represent the decomposed pyrite. The bed is 0.7 m thick.

In a dark central mottle of the bed (Plate II, Fig. 4), there are seen sections of "chervotochinas" representing together with some biomolds one system of pores. Pore walls are locally impregnated by half-decomposed pyrite, as one can judge from sporadic occurrence of cubic crystals in this opaque matter. Pores are locally filled in with serrate sparite and may contain argillans (cutans of clayey composition), which make their shape rounded. The filling-in of both types may indicate the commencement of soil-forming process when plants that left rhizocretions colonized sediments. I failed however to find any sign of subaerial surface above rhizocretions and dark mottles, and the ground exposition above sea level, if it happened, could be very brief, insufficient for the paleosol formation.

Bed 26 (upper coquina; 0.35 m): grayish brown limestone; the bed is cherry in the upper portion, clayey, moderately silicified, very hard, and lacking porosity. Abundant brachiopod shells create the rock framework. The dominant form is *Eomarginifera lobate* (Sow.). Under microscope, the rock corresponds to brachiopod floatstone-rudstone with abundant conodonts and biomolds after molluscan shells (Plate II, Fig. 5). Signs of micritization are rare. Synsedimentary

and early diagenetic voids are filled in with coarse equant sparite. Matrix is partially neomorphic. Compaction signs are lacking.

Member 6 accumulated in a relatively shallow setting favorable for colonization by plants of the mangrove ecotype like those that left behind the third root horizon. Distinct cementation and neomorphism around rhizocretions, as well as the dissolution caverns locally enclosing argillans, point to the short-term events of exsiccation. As beds 22–24 are lacking rhizocretions, those of Bed 21 may mark a separate episode of shoaling. Accordingly, the fourth root horizon may consist of two closely spaced subhorizons. The extreme habitat environments of benthos in response to shoaling resulted in development of monodominant assemblages of micro- and macrofossils.

Member 7 (beds 27–42): Above the basal clay laminae that is conventionally attributed to Member 7, there is dark bituminous packstone (Bed 28) of a shallow-water genesis, as one can judge from mass accumulations of Kamaenidae thalli in the rock. In distinction from underlying member, skeletal remains in Bed 28 and in basal crinoid-brachiopod rudstones of Bed 29 are more diverse and crushed stronger. These features of a higher life activity are characteristic of transgressive facies.

The overlying interval includes black bituminous clay and mudstone-like palygorskite shales with lenses and interlayers of gray dolomitic marl, which are early diagenetic in origin. The impoverished fossil assemblages mostly consist of brachiopods, dominant among which is *E. lobata*. Associate graptolites represent species of the genus *Dictyonema*. The upper half of the member is moderately gypsiferous. The uppermost Bed 42 of lighter-colored marly shale yields another assemblage of more diverse and abundant macrofossils. *Dictyonema* forms are absent here. They gave place to reticulate bryozoans (Fenestellidae), and associated brachiopod assemblage is of a different composition. *Zoophycos* forms dominate among ichnofossil at the base (Bed 28), and small fucoids are widespread higher.

Bed 27 (0.2–0.25 m): brown to dark gray mudstone with accumulations of bioclasts and brachiopod shells near the base (some brachiopods are buried *in situ*). Long crinoid columnals occur as well. Bioclasts decrease in size upward, where their accumulations are absent. Transition to Bed 26 is gradual.

Bed 28 (Tarusa Marble; 0.2 m): dark gray and massive bituminous limestone; the rock is homogeneous and bioturbated. Small (0.2–0.3 cm in diameter) branching fucoids sometimes reveal a transverse sculpturing, but their origin is unclear. Fish detritus is concentrated closer to the base. Visible at the top is gradual transition to black foliated mudstone. *Zoophycos* seems to be a dominant form of ichnofossils. Two textural types are distinguishable under microscope. The first one corresponds to medium-grained bioclastic packstone-wackestone that bears rather abundant foraminifers and spines filled in with sparite (sponge spicules?), shows presence of Kamaenidae occurring as single specimens, and has a considerably dolomitized matrix with cavities of the early generation filled in with coarse-grained sparite (Plate II, Fig. 6). The second variety is algal-brachiopod wackestone-packstone less porous and dolomitized to a lesser extent (Plate II, Fig. 7). In the last case, a considerable part of skeletal remains (more than a half) is represented by Kamaenidae.

Third Bench

Bed 29 (2.6 m): black, very plastic and smeary fissile clay with interlayers and lenses enriched in brachiopod shells, mostly of species *E. lobate* and *Composita ambigua* (Sow.), and also in other macrofossils. Accumulations of graptolites (genus *Dictyonema*) are characteristic of the upper interval. At present, only the upper half of the bed is exposed. According to description of Barskov and Goreva, the bed includes discontinuous interlayers of barite and limonite nodules of a complex cavernous-septate structure, which can be found now in waste piles. The same piles reveal presence of blocks of black brachiopod-crinoidal rudstone (Plate III, Fig. 1) with clayey matrix and abundant brachiopods, among which species *E. lobata* is dominant. This rudstones blocks differ from all rocks visible in exposure and have been likely derived from the lower part of Bed 29.

Beds 30–41 (7.6 m): black clay plastic below and getting fissile upward, especially in beds 39 and 41. Clay grades into dark gray to violet shales, which do not get soaked and are splitting into broad plates resembling cardboard shreds. Rocks of the member are rich in palygorskite (Osipova and Bel'skaya, 1965; Makhlina and Zhulitova, 1984; Makhlina *et al.*, 1993). Clay and shales are intercalated with thick (up to 0.4 m) interlayers and lentils of gray dolomitic marl. Macrofossil assemblages from clay and shales are of a low diversity, consisting predominantly of one–two brachiopod species (*E. lobate* and *Antiquatonia khimenkovi* or large spiriferids of genera *Fusella* and *Unispirifer*). Abundant detritus of fish remains and graptolites of the genus *Dictyonema* are dispersed up to the Bed 41, inclusive. Trace fossils in all rock types (dolomitic marl, clay, and shales) are represented by small abundant fucoids and by rarer *Zoophycos* and *Teichichnus* forms.

Bed 42 (0.4 m): greenish gray palygorskite marl of fissile to flaggy structure; the rock yields abundant macrofossils, predominantly brachiopods. Fenestellid bryozoans resembling *Dictyonema* forms are frequent, but true representatives of this graptolite genus have not been found.

Intercalations of dolomitic marl in Member 7 are especially interesting. Their rocks are of subidiomorphic to idiomorphic texture with clay-filled interstices

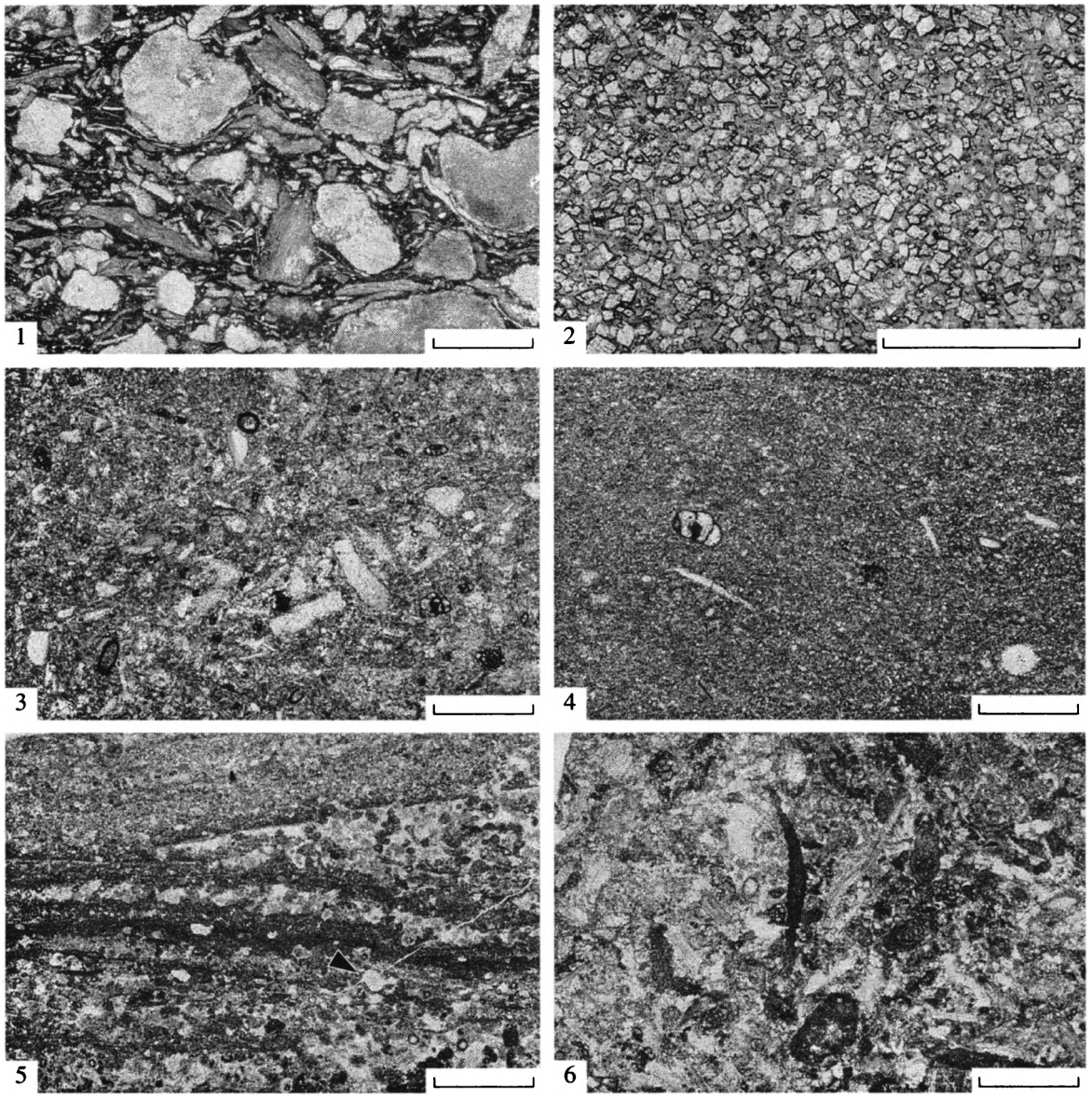

Plate III. Microfacies of upper Steshev and Protva carbonate beds:
(1) A lower portion of Bed 29, brachiopod-crinoidal rudstone (scale bar 1mm). (2) Bed 34, dolomitic marl of a loose idiomorphic crystalline texture; interstices between rhombohedric dolomite crystals are filled in with clay (scale bar 5 mm). (3–6) Microtextures of the *Protva limestones*: (3) a lower portion of Bed 48, bioclastic wackestone; (4) a lower part of Bed 51, mudstone with isolated brachiopod, echinoderm, and foraminifer fragments; matrix in both rocks is recrystallized to the state of microsparite; (5) Bed 53, highly altered microlaminated rock with neomorphic sparite (black arrow) superimposed on the primary texture; (6) Bed 54, intensively leached and recrystallized peloid-bioclastic packstone-grainstone. Scale bars 1mm in figs. 3–6.

between crystals (Plate III, Fig. 2). They preserve corroded conodonts and bioclast traces. The lower dolomitic Bed 30 encloses veins and nests of columnar gypsum. Transitions from dolomitic marl to black clay or shale are always gradual. Some flattened concretion-like bodies of marl have very gradual transitions to dark gray dolomitic shale. Brachiopod casts, *Dictyonema* remains, and bivalves from dolomitic intercalations represent a fossil assemblage similar in composition to though less diverse than that from surrounding clay as a consequence of dolomitization. Macrofossils are flattened in compacted clay and shale beds are almost undistorted in dolomitic marl. It is likely that dolomite is of the early diagenetic origin and does not represent a particular sedimentary fraction. The initially dispersed dolomite crystals and subsequent appearance of dolomitic intercalations were clearly related in origin to the growing aridity of climate, other indications of which were considered earlier (Osipova, Bel'skaya, 1965a, 1977; Makhlina *et al.*, 1993).

Makhlina *et al.*, (1993) suggested to divide the Steshev Horizon (Formation) into the lower (Glazechnino) and upper (Dashkovo) subformations. Member 7 described above includes an upper clayey part of the former and the entire latter subdivision, because the transition from the Glazechnino plastic clay to the Dashkovo palygorskite shales with dolomitic intercalations is gradual and cannot be traced confidently. At the same time, the base and top of Member 7 are distinct and well traceable at least between the Zabor'e and Novogurovskii sections.

Member 7 exhibits a typical sequence of the *Steshev Lagoon* facies from basal black clay with limestone interlayers to upper palygorskite shales with intercalations of dolomitic marl. Based on absence of characteristic representatives of shallow-water benthos, Shvetsov (1932) originally regarded this peculiar sequence as a unit deposited in a deepest setting of the Serpukhovian basin. Later on, he changed his opinion and attributed to shallow-water deposits its upper shaly-dolomitic interval at least (Shvetsov, 1938, 1948). Indications in favor of this interpretation were as follows: the high lateral variability of rock facies, the growing abundance of sandy fraction to the north, the development of dolomite, and exsiccation marks at the top (Member 8, see below). The peculiar composition of fossils was not indicative of either shallow- or deep-water sedimentation environments, but it pointed to considerable paleogeographic changes after the Tarusa time. Osipova and Bel'skaya (1965, 1967, 1977) interpreted the entire clayey sequence described above as deposits of the "Steshev Lagoon," but the lagoonal genesis of its lower beds, at least, seems doubtful. First, the early diagenetic dolomite in upper Moscovian and lower Kasimovian deposits of the same region is typical of rocks that have been formed in a subtidal zone, its deepest part included, but not of sediments characterizing the hyperhaline shoals (Kabanov, 2000). Moreover, muddy sediments of deep-water settings with normal salinity and high concentration of organic matter are very appropriate for dolomitization (Slaughter and Hill, 1991). Second, the peculiar composition of fossils (the impoverished monodominant assemblages of brachiopods, absence or extreme rarity of corals and foraminifers, and appearance of graptolites, small ammonoids and inarticulate brachiopods *Orbiculoidea dalmanensis* Sok., all untypical of the Moscow basin during the Early Carboniferous) can be explained not only by isolation in a lagoon, but also by specific eutrophic and anoxic environments below pycnocline (the water layer with a high density gradient). According to the model of antiestuarine water circulation in epicontinental seas (Witzke, 1987), the necessary stratification of water appears when outflow of dense saline water from near-bottom layer is somehow hindered and the low density of surface water layer is maintained by freshening. Deep-water facies characterizing sedimentation environments of this kind are black phosphorite-bearing clays of the Pennsylvanian cyclothems (Heckel, 1977). Phosphatic nodules present in lower clay beds of the Steshev Horizon (data of A.S. Alekseev, fig. 2) emphasize the analogy of the beds with the mentioned Pennsylvanian clays.

Member 8 (beds 43–46): Limestones of the member are pinkish and intensively silicified near the base and light beige breccia-like in the upper portion. The rocks, which did not experience brecciation, retain sedimentary lamination. Their primary structure was shaded under influence of vadose water during the pre-Moscovian event of subaerial exposition. The mudstone-wackestone texture and lenticular laminae of peloid to oncoid calcarenite are distinguishable under microscope. Skeletal remains are of a uniform and specific composition, represented predominantly by ostracods (Kabanov, in press; Plate I, Figs. 6, 7; Plate II, Fig. 1). Many bioclasts are micritized. The member is crowned by the Dashkovo paleosol.

Bed 43 (0.1 m): pinkish silicified clayey limestone with thin (3–5 mm) laminae and lenses of oncoid calcarenite (Kabanov, in press; Plate I, Fig. 7); visible in thin sections are fragments and intraclasts of microlaminated (stromatolitic) cruststone. Bioturbation signs are invisible. Many clasts are transformed into peloids and rounded. Thin-walled ostracod shells, usually strongly deformed and neomorphic, are frequently distinguishable in oncoid cores.

Bed 44 (0.05 m): clayey aphanitic fissile limestone; the rock is strongly silicified.

Bed 45 (0.8 m): light beige clayey limestone with cherty bands; the rock grades upward into fissile marl. Textural varieties are represented by bioclastic-oncoid mudstone and wackestone with size-variable lenses of oncoid grainstone-rudstone. As a rule, these lenses rest on the erosion surface and grade upward into wackestone; some erosion surfaces are covered by thin microlaminated films with syngenetic drying cracks (Kabanov, in press, Plate II, Fig. 1). Primary interstitial porosity of grainstone-rudstone is cemented by equant sparite. The thickest (0.1–0.15 m) and extended lens of Siphonophycaceae-peloid-oncoid grainstone-rudstone is situated 0.35 m above the bed's base (Kabanov, in press, Plate I, Fig. 6). The gentle indistinct cross bedding is noticeably in the lens. The primary lamination is almost lacking signs of bioturbation, but laminae are frequently deformed and disrupted owing to development, as I think, or drying cracks.

Bed 46 (0.3–0.5 m): yellow ferruginate and silicified limestone; the rock is finely brecciated, corresponding to the calcareous duricrust of the Dashkovo paleosol that will be described later on (Kabanov, in press). Rhizocretions in the duricrust represent the *fifth root horizon.*

The following indications are important for understanding of the Member 8 accumulation conditions. The primary interlamination of wackestone, mudstone, and calcarenite is almost undisturbed by bioturbation. In distinction from lower horizons indicative of shoal-

ing, we observe here calcarenite interlayers overlying the erosion surfaces. The latter could be formed only under influence of high-energy processes. Most considerable grainstone-rudstone interlayers reveal sorting of clasts and inner cross bedding of finer and coarser material. Clasts are usually rounded and associate with predominant oncoids, which formed when periods of immobility alternate with short-term episodes of rolling. Intraclasts of stromatolite-like laminae with margins bent upward are likely indicative of periodic synsedimentary exsiccation episodes, which left behind desiccation mud cracks. Presence of the latter on the member bedding planes was noted by Shvetsov (1948). Macrofossils have not been found, and microfossils are not diverse, mostly represented by ostracods. Siphonophycaceae thalli and biomolds after fragmented molluscan shells are locally abundant. Brachiopod bioclasts are considerably reworked and allochthonous. Echinoderm remains abundant below are missing from the member. So it is likely that Member 8 corresponds to the facies of a low-energy tidal plain (predominance of wackestone and mudstone beds) that was influenced sometimes by high-energy events in distinction from shoaling episodes recorded below. Like in present-day tidal plains (Aigner, 1985; Wanless *et al.*, 1988; Tucker, Wright, 1990), storms were responsible for these events. The peculiar composition of sedimentary grains, which differ from material deposited earlier, suggests that the plain in question was well isolated from the main basin of sedimentation.

PROTVA HORIZON

The results obtained clarify the character and position of the boundary between the Steshev and Protva horizons. Shvetsov (1940, p. 274) defined this boundary in the region at the level "of disappearance of abundant and diverse fauna, above which microlaminated rocks (clay and chemogenic limestone overlying it) are either barren, or overfull with ostracods, while other fossils are absent." Afterward (Shvetsov, 1948, p. 138), he pointed out that the "upper member of flaggy chemogenic limestones" is well traceable in the west of the southern limb of the Moscow syneclise, the Serpukhov area included. This unit is easily recognizable in the described Member 8 that is above the boundary between the Steshev and Protva horizons as it is accepted in some works (Makhlina *et al.*,1993; Barskov and Alekseev, 1979; Skompsky *et al.*, 1995). It seems probable that Shvetsov included in the interval of barren chemogenic limestones the calcrete duricrust of Dashkovo paleosol that had not been recognized at that time. Thus, it is reasonable to place boundary between the Steshev and Protva horizons slightly higher, at the level of transition from shales to limestones, i.e., at the top of the Dashkovo paleosol that crowns the member of chemogenic limestones with ostracodes and represents the real boundary between the "sixth and seventh Dinantian cyclothems" in terminology of Shvetsov (1938).

Member 9 (beds 47–51): The member is composed of grayish to white massive limestones, which are karstified, recrystallized, stained with iron compounds, and rest on a basal layer of palygorskite clay. Packstone and wackestone beds 48 and 49 are intensively bioturbated (*Zoophycos* structures among others) and yield an assemblage of large and small skeletal remains of normal marine affinity. Coarse-grained storm concentrates occur as well. In the upper beds 50 and 51 of bioclastic wackestone and mudstone, remains of small skeletal fossils are slightly less diverse. The member is 2.15 m thick.

Bed 47 (0.1–0.15 m): greenish brown palygorskite clay; the rock is fissile, yielding large ostracods.

Bed 48 (0.55 m): indistinctly granular to aphanitic, relatively monolithic limestone with rusty stains and *Zoophycos* structures; the rock yields mollusks, brachiopods, and other macrofossils. Calcite druses related in origin with the pre-Moscovian karstification are characteristic of the upper part. Near the top, there are subvertical twisting fucoids. The topmost portion is composed of foliated clay laminae. Textural varieties distinguishable under microscope are fine-grained, bioclastic, intensively bioturbated packstone and wackestone (Plate III, Fig. 3).

Bed 49 (0.15 m): slightly clayey compact limestone with *Zoophycos* traces and large abundant bellerofontids; under microscope, this is bioclastic, predominantly brachiopod wackestone-packstone. Skeletal fragments are diverse and irregularly distributed. Larger clasts (>1–2 mm) form segregations and lenses with the rudstone texture, which are depleted in micritic material and separated from each other by areas of clayey wackestone. The concentrations probably exemplfy remnants of basal graded storm concentrates, which have been disintegrated by bioturbation.

Bed 50 (0.1 m): greenish gray clayey and hard aphanitic flaggy limestone; an almost continuous chain of black chert nodules replace locally the entire bed's thickness.

Bed 51 (1.2 m): aphanitic limestone consisting of alternating monolithic and nodular interlayers; under microscope, textural varieties correspond to mudstone or ostracod-brachiopod wackestone (Plate III, Fig. 4) and to floatstone in places of concentrated shells.

The palygorskite clay of Bed 47 with large ostracods was likely deposited during a quick transgression (insignificant bed's thickness) that flooded the Dashkovo paleosol and created normal marine conditions in a sublittoral zone, where the overlying beds accumulated. Beds 48 and 49 with diverse skeletal remains, *Zoophycos* traces, and bioturbation structures enclose remnants of storm deposits and correspond to a period of normal marine sedimentation in the subtidal zone that experienced episodic impacts of stormy waves. Microscopic indications of shallow-water environments, such as tubular algae and micritization, are extremely rare in pure limestones, and the depth at the

time of their deposition should be not less than few dozens of meters. The transition to upper muddy lithofacies indicates the hydrodynamic activity lowering in sedimentation settings, probably in response to commencement of shoaling.

Member 10 (beds 52–54). The member is composed of highly karstified marls and limestones whose sequence rests on a basal clay bed. In its original stratification, the member is exposed in several areas of the eastern quarry wall. The karst breccia exposed between these areas is buried in the Vysokoe Clay. Flaggy marl and clay deposited above the basal Bed 52 locally reveal microlamination. The terminal part of the member consists of shallow-water bioclastic limestones with chaetetids. The upper boundary of the member is conventionally placed at the top of southern outlier in the eastern quarry wall, limestones above which are preserved in a form of silicified breccia. The member is about 1.7 m thick.

Bed 52 (0.3 m): yellow ferruginate plastic clay; white calcareous flour seen locally near the top is likely related in origin with development of the pre-Moscovian weathering crust.

Bed 53 (about 0.3–0.4 m): greenish yellow marl and clay; these rocks of flaggy to foliated structure are silicified and locally transformed into banded chert. Gentle wavy microlamination visible in a single thin section exemplifies alternation of fine-grained peloid calcarenite and mudstone laminae up to 1–3 mm in thickness (Plate III, Fig. 5). Rosette-like inclusions of coarse-crystalline sparite are clearly visible against the background of recrystallized primary texture.

Bed 54 (about 1.0 m): light yellow thick-bedded limestone; the rock indistinctly granular below grades upward into the aphanitic variety and bears chaetetids. A thin section from the lower granular portion exemplifies peloid-bioclastic bioturbated packstone-grainstone, many clasts of which are micritized (Plate III, Fig. 6).

Beds 52–54 are conventionally attributed to a single member, because their rocks are clearly different visually and under microscope. Owing to the intense karstification located between the beds, it is unclear whether they are interconnected via gradual transitions or separated by unconformities. At any rate, the diverse skeletal benthos and *Zoophycos* traces, which could point to sedimentation in normal marine subtidal settings, have not been distinguished in the beds. Chaetetids may occur in both the normal marine and marginal facies. Two thin sections from beds 53 and 54 show that their rocks originated in shallower settings than those of beds 48 and 49. In general, it is possible to assume the shallow-water origin of the entire Member 10 in contrast to the rocks of Member 9, which clearly were deposited in deeper environments.

UPPER LOWER CARBONIFEROUS AND BASHKIRIAN STAGE OF THE MIDDLE CARBONIFEROUS

Vysokoe Beds

Bed 55: silty clay with limestone breccia; together with the karstified limestone of the Protva Horizon, the rocks are attributed to the pre-Moscovian weathering profile that will be described elsewhere (Kabanov, in press).

Bed 55 crowns the Carboniferous rock sequence exposed in the quarry. The overlying Quaternary deposits include a cultural level.

CONCLUSION

Investigation results obtained for the Zabor'e section agree in general with and simultaneously elucidate some new details of the Serpukhovian Stage structure previously established in the stratotype area.

(1) The section is divided into ten members corresponding to consecutive stages in development of the regional paleobasin. The basal Member 1 (1.3 m) of karstified bioclastic wackestones and packstones is terminal one in the Venev Horizon. Members 2 to 6 characterizing the lower 13-m-thick interval of the Serpukhovian Stage correspond to the Tarusa Horizon and to the lower half of the Steshev Horizon. Rocks of the former are represented by gray bioturbated micrite-bioclastic limestones, which are overlain by the lower Steshev clayey limestone and marl beds with clay intercalations. The established microfacies correspond to bioclastic wackestone, packstone, and floatstone. *Zoophycos* forms distinctly prevail over others in the ichnocoenosis. Rounded clasts and high-energy facies are almost absent, except for infrequent intercalations of storm sediments, which appear near the base of the Steshev Horizon. Member 7 (8.15 m) or facies of the "Steshev Lagoon" are represented by black plastic clay that yields an oligodominant assemblage of macrofossils. The clay grades upward into flaggy palygorskite shales with intercalations of early diagenetic marl. The upper 6-m-thick interval of the section (upper Steshev and lower Protva beds) is composed of light-colored limestones with marl and clay interlayers, which had been intensively altered at the time of the pre-Moscovian weathering episode. Member 8 (1.25 m) crowning the Steshev Horizon is represented by oncoid-ostracod laminite with recurrent marl and limestone laminae, and by calcrete at the top. The Protva Horizon interval (4.85 m) includes Member 9 of bioturbated sediments with *Zoophycos* forms (relatively deep-water mudstones, wackestones, and packstones) and Member 10 of highly weathered flaggy marls and micritic to peloid-bioclastic limestones, which seem to be of a shallower genesis.

(2) Four to five levels of shoaling and three profiles of subaerial exposition, all formerly unknown, are distinguished in the section. The latter will be described in

a subsequent paper (Kabanov, in press). Member 3 has a hummocky upper surface that most likely represented a hardground, though it could be of subaerial origin as well. The confident shoaling levels bear rhizocretions of presumably mangrove plants and exhibit marks of minor subaerial alterations. Above the *first root horizon* corresponding to the Muratovo paleokarst, the shoaling levels in the lower third and near the top of the Tarusa Horizon and in the middle of the Steshev Horizon are represented by the *second, third, and fourth root horizons*, respectively. Surfaces of subaerial expositions do not occur above these levels. Member 4 (2.3 m) enclosing the third root horizon is distinctive because of transition to packstone with abundant algal remains of the Kamaenidae family, partially preserved microlamination, and absence of *Zoophycos* forms. Two other levels are lacking considerable changes in biofacies. The fourth level may correspond to a couple of closely spaced root horizons. The *fifth root horizon* includes rhizocretions of the Dashkovo paleosol. It is likely that shoaling levels can be found as well in the uppermost, highly altered limestone beds of the Protva Horizon.

(3) Following the cyclostratigraphic principle of defining boundaries of local and regional subdivisions, it seems reasonable to place the Tarusa–Steshev boundary at the level of maximum shoaling inside Member 4, i.e., 1.1 m below that placed formerly at the base of Member 9. The Steshev–Protva boundary is suggested to be at the top of the Dashkovo paleosol, the level of which is 1.25 to 1.65 m higher than that of formerly accepted boundary.

(4) The earlier assumed lagoonal origin of the Steshev clayey sequence (Osipova and Bel'skaya, 1965a) seems doubtful for its lower portion at least. In the alternative model suggested here, the bituminous clay beds of the sequence are interpreted as sediments accumulated below pycnocline and wave base level in a basin with density-stratified water column.

(5) The origin of graded storm deposits persistently appearing near the Steshev Horizon base in the Zabor'e and Novogurovskii sections cannot be explained by the facies variability. In the terminal portion of the horizon, there are intertidal laminites with indications of storm impact. If the appearance level of storm deposits could be traceable in other sections of the study region, then, taking into account the indications of aridity growth during the Serpukhovian Age (Osipova and Bel'skaya, 1977), we may characterize the regional climatic changes as a transition from the humid wind-free climate of the Okaian–Tarusa time to the subsequent period of arid climate and enhanced atmospheric circulation that resulted in development of storm events. It is also probable that paleoclimatic changes recorded in the section reflect development of the Gondwana Glaciation, glacioeustatic control of sea-level fluctuations, and coeval climatic differentiation of the Earth.

ACKNOWLEDGMENTS

I am grateful to my colleagues N.B. Gibshman, I.S. Barskov, S.V. Nikolaeva, A.S. Alekseev, and B.T. Yanin for collaboration in the Zabor'e Quarry, fruitful discussions, and constructive comments to this and other manuscripts concerning the issue. I would like to thank M.K. Emel'yanova for the technical assistance. The work was supported by the Russian Foundation for Basic Research, project nos. 99-05-65473 and 00-15-97-764.

Reviewers P.Yu. Petrov and N.M. Chumakov

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Foraminifers from the Serpukhovian Stage Stratotype, the Zabor'e Quarry Site (Moscow region)

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Received October 26, 2000

Abstract—Distribution of foraminifers in the Tarusa, Steshev, and Protva horizons of the Serpukhovian Stage stratotype has been studied in the Zabor'e Quarry. Boundaries of *Pseudoendothyra globosa, Eostaffellina decurta*, and *Eostaffellina protvae* zones distinguished here are defined at the first occurrence levels of zonal index species. As is established, boundaries between horizons and foraminiferal zones do not coincide. *Neoarchaediscus postrugosus* (Reitl.) and "*Millerella*" *tortula* Zeller, which appear at the boundary between the Visean and Serpukhovian stages, can be regarded as markers of that stratigraphic level. Species "*Millerella*" *tortula* Zeller, *Janischewskina delicata* Mal., and *Brenckleina rugosa* (Brazhn.), which are important for stratigraphic subdivision and correlation, have been identified for the first time in the Moscow region. It is shown based on the microfacies analysis that foraminifers populated calm settings of a shallow sea that was shoaling with time up to the appearance of algal banks.

Key words: Serpukhovian Stage; Venev, Tarusa, Steshev, and Protva, horizons, foraminiferal zonation, microfacies, bioclasts.

INTRODUCTION

The Serpukhovian Stage first distinguished in Russia (Nikitin, 1890, Shvetsov, 1922, 1948) is a possible candidate for the standard subdivision of the Carboniferous System. This perspective, highly dependent on the progress in comprehensive investigations of the type sections in Moscow region, stimulates interest to searching here a complete section that can be subdivided into horizons and biozones based on different groups of paleontological remains and established appearance levels of index species applicable for correlation.

Nikitin (1890) regarded the Zabor'e Quarry, Luzhki, and Podmokloe sections as most typical of the Serpukhovian Stage (Fig. 1). The first one is at present in the status of stratotype and exhibits Serpukhovian deposits accessible for investigation (Rauser-Chernousova, 1948a; Osipova and Bel'skaya 1965, Makhlina *et al.*, 1993; Skompsky *et al.*, 1995). Above the upper Visean Venev Horizon, there are exposed the Tarusa, Steshev, and Protva horizons. The section was divided into foraminiferal and conodont zones by Makhlina *et al.* (1993).

INITIAL INVESTIGATIONS OF FORAMINIFERS FROM THE ZABOR'E QUARRY

An incomplete geological column of the Zabor'e Quarry that has been published by Rauser-Chernousova (1948a) includes the Tarusa and Steshev horizons. Flaggy limestones exposed in the lower part of the quarry are attributed to the Tarusa Horizon (beds 1–4), and overlying lower strata of dark clay intercalated with flaggy limestones partially represent, together with the upper fissile clay beds, the Steshev Horizon (beds 5–7). The bed-by-bed distribution of foraminifers collected by Rauser-Chernousova is shown in Fig. 1.

Among foraminifers from the Tarusa Horizon studied in detail, she identified *Earlandia vulgaris* (Rauser et Reitl.), *E. minor* (Rauser et Reitl.) (given with a question mark), *Endothyra bradyi* Mikh. (frequent or common), *Endothyranopsis sphaerica* (Rauser et Reitl.) (frequent), *Archaediscus* cf. *karreri* Brady, *A.* cf. *krestovnikovi* Rauser (at the base of the horizon), *A. krestovnikovi* Rauser (usual near the horizon top), *Neoarchaediscus* cf. ex gr. *rugosus* (Rauser) (in the middle part), *Asteroarchaediscus ovoides* (Rauser) (frequent in the middle part), *Endostaffella parva* (Moeller) (dispersed throughout the horizon), *Mediocris* cf. *mediocris* (Viss.) (single in the middle part), *Eostaffella* cf. *mosquensis* Viss. (single in the middle part), and *E. parastruvei* Rauser (frequent throughout the section). All the listed species of wide stratigraphic ranges are inherited from the upper Visean Substage. Unfortunately, *Parastaffella struvei serpukhovi* that was identified as a new subspecies has not been described and pictured afterward.

In the Steshev Horizon that was studied from its lower boundary to the middle (beds 5–7), there were identified species *Endothyra bradyi* Mikh. (the horizon base), *Archaediscus karreri* Brady (the horizon base), and *A.* sp. (middle part), which are inherited from the

Tarusa Horizon. They associate with newcomers *A. pusillus* Rauser (frequent), *Asteroarchaediscus.* cf. *bashkiricus* (Krest. et Theod.), and *Neoarchaediscus parvus* (Rauser) (frequent), all having first occurrence levels at the horizon base.

The boundary between Tarusa and Steshev horizons was placed at the base of dark clay in accordance with the lithological transition and following the opinion of Shvetsov (1922, 1948) widely acknowledged at that time. The upper section part, formerly unexposed and corresponding to the Steshev and Protva horizons, had not been investigated.

Thus, the foraminifer assemblage from the lower portion of the Serpukhovian Stage stratotype turned out to be impoverished, mostly consisting of transit late Visean species.

Recently, M.V. Vdovenko (see in Makhlina *et al.*, 1993) reinvestigated foraminifers from the Zabor'e Quarry and defined for the first time the foraminiferal zones in this section, which are accepted in the regional stratigraphic scheme for the Russian platform (*Reshenie...*, 1990). Vdovenko approved data of Rauser-Chernousova (1948a) on foraminiferal assemblage from the Tarusa Horizon and added new taxa to the former list of foraminifers from the Steshev Horizon. She also pointed to some foraminiferal taxa present in the Protva Horizon.

According to Vdovenko, the *Pseudoendothyra globosa–Neoarchaediscus parvus* Zone (beds 3–10) characterizes the Tarusa (one assemblage from beds 3 and 4) and Steshev (three assemblages from beds 5–6, 7–8, and 9–10) horizons. Her list of foraminifers (appendix 3 in the mentioned work) includes all taxa identified by Rauser-Chernousova (1948a), except for *Endothyranopsis sphaerica* (Rauser et Reitl.), and newly identified forms. The latter are *Earlandia minima* (Bir.), *E. elegans* (Rauser et Reitl.), *Palaeotextularia* sp., *Endothyra* sp., *E.* ex gr. *phrissa* (Zeller), *E. similes* Rauser et Reitl., *Omphalotis* sp., *Endothyranopsis crassa* (Brady), *Betpakodiscus (?)* ex gr. *compressa* (Vdov.), *Tetrataxis eominima* Viss., *Endostaffella delicata* Ros., *E.* ex gr. *asymmetrica* Ros., *E. fucoids* Ros., *Euxinita* sp., *Loeblichia ukrainica* Brazhn., *Quasiammodiscus buskensis* (Brazhn.), *Planoendothyra* sp., *Eostaffella* ex gr. *tenebrosa* Viss., *Eostaffella* ex gr. *postmosquensis* Kir., *Plectostaffella* aff. *varvariensis* (Brazhn. et Pot.), and *P.* ex gr. *varvariensis* (Brazhn. et Pot.). Vdovenko failed to detect the index species *Pseudoendothyra globosa* Ros. and to note changes in foraminiferal assemblages at the boundary between horizons. Interesting among forms supplementing list by Rauser-Chernousova (1948a) are *E.* ex gr. *phrissa* (Zeller), *Planoendothyra* sp., and *Betpakodiscus* (?) ex gr. *compressa* (Vdov.) unknown from the upper Visean beds. In addition, *Eostaffella* ex gr. *postmosquensis* Kir., *Plectostaffella* aff. *varvariensis* (Brazhn. et Pot.), and *P.* ex gr. *varvariensis* (Brazhn. et Pot.) are younger than others, because species *Eostaffella postmosquensis* Kir. (holotype from the Kamenka site at the Chusovaya River) has been described (Rauser-Chernousova *et al.*,1951) from deposits of the Bashkirian Stage (Podvereiskii Horizon), being unknown below the upper Serpukhovian Substage. In addition, species *Plectostaffella varvariensis* (Brazhn. et Pot.) were originally described from Limestone D-4 (Brazhnikova *et al.*, 1948), and representatives of this group do not occur below the Zapaltyube Horizon in the Donetsk basin (Aizenverg *et al.*, 1983) and other regions (Kulagina, 1988; Kulagina *et al.*, 1992; Gibshman *et al.*, 1990).

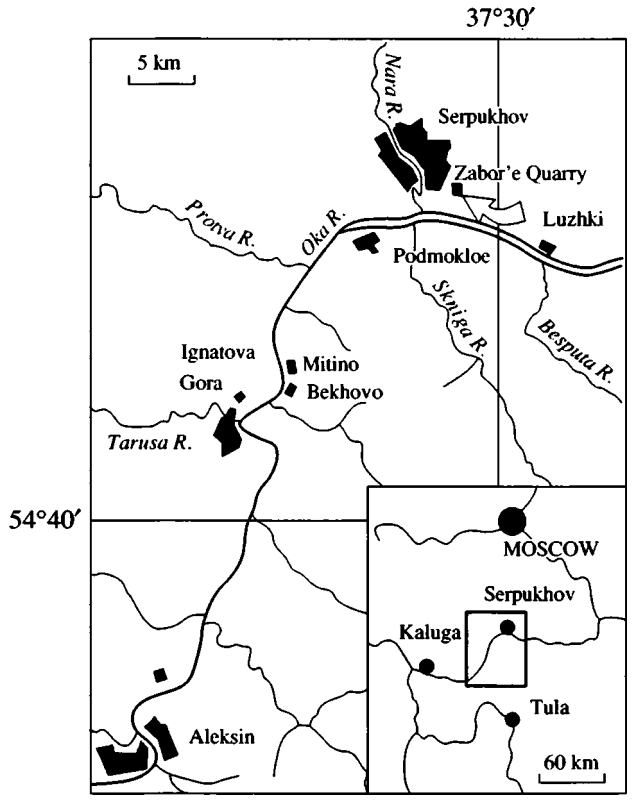


Fig. 1. Geographic sketch map of Moscow region with the indicated Zabor'e Quarry and other localities of the type Serpukhovian sequences.

Vdovenko established the *Eostaffellina protvae* Zone in the lower part of the Protva Horizon, where she identified *Parastaffella angulata* Rauser, *Pseudoammodiscus* sp., *Endothyra* sp., and *Eostaffellina* sp.

Thus, even after results obtained by Vdovenko, it seemed that foraminifers from the Serpukhovian Stage stratotype are mostly represented by transit Visean forms, whereas stratigraphically important, typical Serpukhovian taxa are missing among them, except for *Neoarchaediscus parvus* (Rauser), *N.* ex gr. *rugosus* (Rauser), *Endothyra* ex gr. *phrissa* (Zeller), *Planoendothyra* sp., and *Eostaffellina* sp. In the meantime, disappearance of late Visean (Okaian) foraminifers is a main criterion of the Serpukhovian Stage discrimination that was suggested long ago, when ranking the

Table 1. Approximate coordination between different systems of bed numeration

Horizon	Rauser-Chernousova, 1948a	Makhlina *et al.*, 1993	Barskov and Goreva, 1978; Barskov, 1999 (numbered samples are those studied for conodonts, Makhlina *et al.*, 1993)
Protva	unstudied	beds 14 and 15, samples 29–35	beds 48–53, samples 42–51
Steshev	beds 5–7, samples 281–292	beds 5–13* (lower part), samples 7–27	beds 10–47, samples 7p–40
Tarusa	beds 1–4, samples 274–280	beds 3 and 4, samples 1–6; 16. beds 1–9	samples 2c and 6b
Venev	unstudied	unstudied	unstudied

*Bed 13 (=Beds 11–13, uniform in lithology).

stage as an independent subdivision (Rauser-Chernousova *et al.*, 1936, a.162), and approved by investigation of sections in the Samarskaya Luka area, southern limb of Moscow syneclise, central Russian platform, and Cis-Urals (Rauser-Chernousova *et al.*, 1940, Rauser-Chernousova 1948 a, b).

In contrast to eastern sections of the Russian platform (Fig. 1), which yielded diverse foraminifers of the Serpukhovian Stage, the assemblage from the Zabor'e Quarry looked extremely impoverished. Besides the initial list of foraminiferal taxa (Rauser-Chernousova, 1948a), only 13, not very important forms were pictured later on (Makhlina *et al.*, 1993).

Like foraminifers, conodonts found in the Zabor'e section (Skompsky *et al.*, 1995) did not improve the correlation potential of the section accepted for the Serpukhovian Stage lectostratotype. Brachiopods, the main paleontological group used by the initial recognition of horizons in the Serpukhovian Stage (Shvetsov, 1922, 1932), are poorly studied in the section. It is known only that "... brachiopods are scarce, mostly represented by spiriferids and smaller productids, whereas giant forms of the latter occur as single specimens" (Osipova and Bel'skaya 1965, p. 7).

Moreover, lithological facies of the Serpukhovian Stage in the Zabor'e Quarry are unknown in detail and have been discussed in general only (Osipova and Bel'skaya 1965) by comparison with results obtained by Shvetsov (1932, 1948) for the Tarusa Horizon of the Moscow basin. According to data of Osipova and Bel'skaya, the Tarusa Horizon of Zabor'e section differs from coeval subdivisions of other sections (the detailed lithology of the latter has not been illustrated and described; section names are missing in the map) and shows predominance of the first-type limestones (in terminology of Shvetsov and Birina, 1935). As they mention in the text, the horizon yields "... diverse Coelenterata (chaetetids, branching and solitary rugoses), abundant nautiloids, and scarce brachiopods. One bed is rich in pelecypod remains (with reference to Astaf'eva-Urbaitis); gastropods, large crinoid columnals, fish teeth, and fucoids (vertical cylindrical and *Thaonurus*) are present as well. Limestones of this level are very similar to those of the Venev Horizon, and their lower beds also bear abundant algae remains *Calcifolium okense*..." Subsequent works on foraminifers from the Serpukhovian Stage of Moscow region (Fomina, 1969, 1977; Osipova and Bel'skaya 1975; Makhlina and Zhulitova, 1984) are lacking descriptions of the Zabor'e section.

The complete geological column with indicated beds and carbonate rock types, which are exposed in the Zabor'e Quarry, has been published without description (Makhlina *et al.*, 1993). In the text (pp. 99 and 102), the section is mentioned among others studied in the paper part devoted to characterization of the Tarusa Horizon.

Still there was a hope that a more comprehensive examination may result in discovery of diverse foraminiferal assemblages from the Serpukhovian Stage stratotype.

Materials for this work have been collected in two periods. In August 1998, during the International Excursion to stratotype sections of Moscow region, I sampled beds 3–6, 15, 16, 18–21, 28, 30, 34, 48, and 49. Preliminary examined thin sections demonstrated a considerable diversity of foraminifers, among which zonal index species useful for interregional correlation were recognized (Gibshman, 1999).

The whole section was sampled once again in June 2000. In both cases, I used bed numbers after Barskov and Goreva (1978), which were accepted by Makhlina *et al.*, (1993) when they studied conodonts, and repeated by Barskov in 1999 almost without changes. Coordination between bed numbers used in different periods (Rauser-Chernousova, 1948a; Makhlina *et al.*, 1993) is shown in Table 1.

In this work, I studied foraminifers in 400 thin sections characterizing 39 rock samples; 350 of them are from own collection and others from collection of P.B. Kabanov. From 10 to 33 thin sections per a unit characterize beds 3–28, 48, and 49, and other beds are exemplified by 5 thin sections per each. The terrigenous parts of the sections (beds 28, 29, and 36–42) have not been sampled. Beds 30, 32, 34, and 36 turned out to be lacking foraminifers. Foraminifers from Bed 2 are

identified in 6 large thin sections from collection of Kabanov, and other materials represent my own collection. The collection is stored at the Paleontological Institute, Russian Academy of Sciences.

All species are identified in accord with the current taxonomy of Paleozoic foraminifers (Vdovenko *et al.*, 1993; Rauser-Chernousova *et al.*, 1996). However, index species of *Eostaffella decurta* and *Eostaffella protvae* zones described below represent an exception, because I had to retain the original zone names introduced by Lipina and Reitlinger (1970), while the subgenus *Eostaffellina* Reitlinger, 1963, was transposed into the genus rank after their publication (Rauser-Chernousova *et al.*, 1996). In description of microfacies and reconstructed sedimentation settings, I used classification of carbonate rocks suggested by Danham (1962) with some additions after Wilson (1975).

STRUCTURE, MICROFACIES, AND FORAMINIFERAL ASSEMBLAGES OF THE SECTION

Rocks of the Zabor'e Quarry in southern outskirts of Serpukhov (Fig. 1) are perfectly exposed. Their sequence is composed of limestones, dolostones, marls, and mudstones. Limestones are dominant rock types in the greater lower and upper parts of the section, and other rocks indicated above are concentrated in its lesser middle interval.

The Serpukhovian Stage was divided into horizons based on lithological criteria and changes in composition of brachiopod assemblages (Shvetsov, 1922, 1932; Shvetsov *et al.*, 1935). The Tarusa Horizon (stratotype near the village of Bekhovo upstream of Serpukhov along the Oka River; Shvetsov, 1932) was distinguished above the top of rhizoid limestone of the Venev Horizon, which marked the Visean–Serpukhovian boundary. The Tarusa Horizon predominantly consisted of beds of compact flaggy limestone in distinction from facies-variable deposits of the Venev Horizon. Its characteristic brachiopod assemblage included abundant *Striatifera striata* Fisch.

The Steshev Horizon (stratotype at the synonymous rapids in upper courses of the Volga River; Khimenkov,1934) was of a more heterogeneous lithological composition. Its lower part corresponded to the so-called Horizon of Crinoid Limestone and "Lobatus" Mudstone, the latter bearing brachiopods resembling in morphology *Eomarginifera lobate* Sow., and the upper part of different lithological composition was known as "Horizon of Fissile Mudstone."

The Protva Horizon of light-colored compact limestones, locally dolomitized and intercalated with cherty lenticular interbeds, and basal layers of variegated mudstone was also known as "Horizon of Sugary Limestones." It was named after the eponymous river, along which exposures of corresponding rocks are most widespread.

The most complete section of the Protva Horizon is situated near the village of Kremenskoe at the Luzha River, where holotype of *Eostaffellina protvae* (Rauser) was found. Geological column and bed-by-bed lithological description of its rocks was published by Osipova and Bel'skaya (1975, pp. 58, 171) who limited its paleontological characterization by a general list of taxa.

In this work devoted to foraminifers, their host rocks are simply named, and main attention is paid to characterization of microfacies. The section is described from its base upward, i.e., from older to younger deposits.

UPPER VISEAN SUBSTAGE, VENEV HORIZON

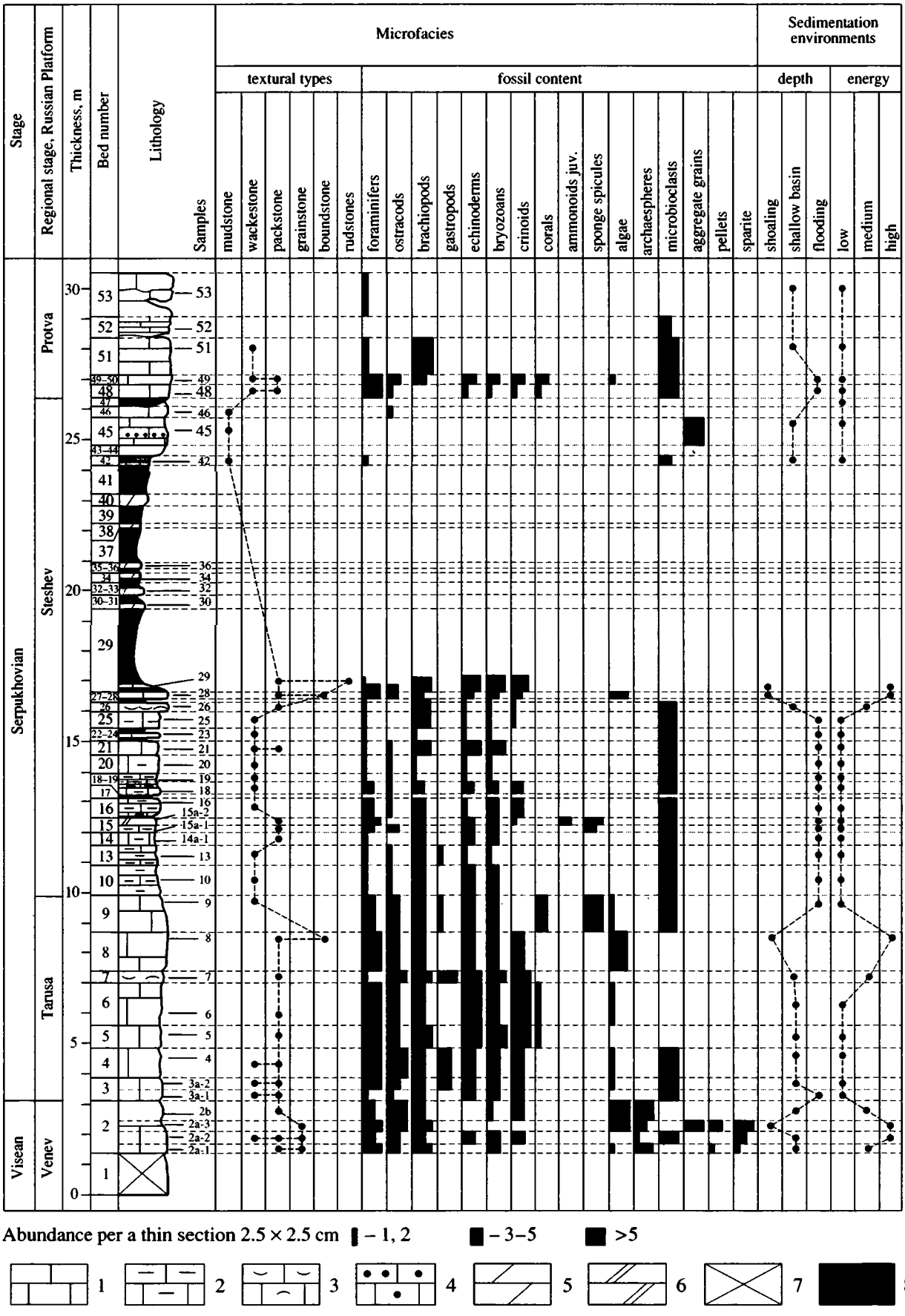
Bed 1 is now concealed under water and has not been studied

Bed 2 consisting of two lithological members has been sampled at four levels (Fig. 2).

Its basal part that is exposed (Sample 2a-1): nearly white bioclastic packstone with minute caverns; the rock is of brachiopod–foraminiferal type, grading into grainstone with pellets (0.021–0.035 mm). Organic remains are represented by brachiopod fragments, rare bryozoans and large ostracods, single specimens of algae *Calcifolium okense* Schwetz. et Bir., and by frequent archaespheres. Foraminifer taxa (20) are of a common abundance and diversity at this level. Their assemblage consists of taxa typical of the upper Visean Horizon and *Eostaffella tenebrosa–Endothyranopsis sphaerica* Zone. It includes rare *Asteroarchaediscus bashkiricus* (Krest. et Theod.) and single *Neoarchaediscus* ex gr. *rugosus* (Rauser) of a narrow shape (Table 2). *Endostaffella* forms distinctly prevail over others.

Middle part (Sample 2a-2): bioclastic wackestone grading into microbioclastic packstone and locally into grainstone. In packstone, abundant microbioclasts are set in a homogeneous sparite matrix, the crystalline structure of which is indistinct. Organic remains are rare, represented by fragmented brachiopod shells, ostracods, bivalves, and single archaespheres in association with powdered skeletons of unidentifiable crinoids or echinoderms. Foraminifers (13 taxa) are less diverse than below. Of two zonal index species, only *Endothothyranopsis sphaerica* (Rauser et Reitl.) is present. Associated taxa are *Bradyina rotula* Eichw., *Mikhailovella gracilis* (Rauser), and *Loeblichia paraammonoides* Brazhn.

Top portion (Sample 2a-3): gray medium-grained grainstone of peloid–bioclastic type. Sparite matrix is distinctly crystalline, bearing infrequent pellets (0.021–0.035 mm) and loose ooze grains. Bioclasts of medium size are represented mostly by fragments of brachiopods, large ostracod shells, abundant algae *Calcifolium okense* Bir. et Schwetz., and archaespheres. The more diverse foraminiferal assemblage includes 16 moderately preserved species, many tests of which are broken.


Stage
Regional stage, Russian Platform
Thickness, m
Bed number
Lithology
Samples
Microfacies
textural types
fossil content
Sedimentation environments
depth
energy
mudstone
wackestone
packstone
grainstone
boundstone
rudstones
foraminifers
ostracods
brachiopods
gastropods
echinoderms
bryozoans
crinoids
corals
ammonoids juv.
sponge spicules
algae
archaespheres
microbioclasts
aggregate grains
pellets
sparite
shoaling
shallow basin
flooding
low
medium
high
Serpukhovian
Visean
Protva
Steshev
Tarusa
Venev
Abundance per a thin section 2.5 × 2.5 cm – 1, 2 – 3–5 >5
1
2
3
4
5
6
7
8

The assemblage includes both index species, though tests of *Eostaffella tenebrosa*? Viss. are scarce and broken. The species associate with taxa typical of the upper Visean Venev Horizon. These are *Globoendothyra globula* (Eichw.), *Endothyranopsis sphaerica* (Rauser et Reitl.), *Omphalotis omphalotis* (Rauser et Reitl.), *Janischewskina* cf. *typical* Mikh., *"Eostaffella" parastruvei* Rauser, *Parastaffella propinqua* Viss., and *P.* spp. cf. *Climacammina* sp. Many forms are frequent. Dominant are *Janischewskina* cf. *typical* Mikh., *Bradyina rotula* Eichw., *Globoendothyra globula* (Brady), and *Archaediscus gigas* Rauser (Table 2), all of a wide range. The euryfacies calm-water representatives, like *Eostaffella, Endostaffella, Asteroarchaediscus*, and *Neoarchaediscus* forms, are absent.

Sample 2b: unsorted bioclastic packstone with clasts of variable size; the rock is black owing to abundant organic matter. Fossils are represented by frequent and large ostracod shells, brachiopod fragments, rare crinoid columnals, and bryozoan remains. Algae *Calcifolium okense* Bir. et Schwetz. and spheroids are much more frequent at this level than below, whereas foraminifers are less abundant and diverse (12 species). Principal components of foraminiferal assemblage are *Endostaffella* and *Archaediscus* forms, whereas *Eostaffella tenebrosa* Viss. occurs as single specimens and *Endothyranopsis sphaerica* (Rauser et Reitl) has not been encountered (Table 2).

The apparent thickness of Bed 2 is about 0.6 m (hereinafter data of Kabanov).

The bed's top is accepted for the boundary between the Visean and Serpukhovian stages. The top portion corresponds to rhizoid limestone crowning the Venev Horizon (Shvetsov, 1932).

SERPUKHOVIAN STAGE, TARUSA HORIZON

Bed 3 is sampled at two levels.

Basal interval (Sample 3a-1): light-colored soft fine-grained limestone of a type transitional between packstone and wackestone; rare areas of amorphous sparite in the rock enclose large bioclasts, which are free of micritization and show transportation marks. Identified among bioclasts are remains of brachiopods, crinoids, echinoderms, and rare ostracods.

Foraminifers occur as single, hardly identifiable specimens of low diversity (6 taxa). Smaller, poorly identifiable *Eostaffella* and *Endostaffella* forms (10 tests) are dominant among them. Other encountered species are *Endothyranopsis* cf. *sphaerica* (Rauser et Reitl.) (one test), *"Endostaffella" asymmetrica* Ros. (2 tests), and *Eostaffella tenebrosa(?)* Viss. (oblique section of one test).

Top interval (Sample 3a-2): fine- to medium-grained wackestone–packstone similar to the previous rock. Fragmented remains of echinoderms, brachiopods, gastropods, pelecypods (?), crinoids, bryozoans, ostracods, and Dasycladaceae (green algae) are set in the microbioclastic matrix. Observable proportions between different types of bioclastic material are approximately equal.

Foraminifers (Table 2) are frequent and rather diverse (18 species). Their predominant forms are *Archaediscus krestovnikovi* Rauser, *A. nanus* Rauser, *Asteroarchaediscus bashkiricus* (Krest. et Theod.), and *Neoarchaediscus* ex gr. *rugosus* (Rauser). Rare specimens represent *Endothyranopsis* ex gr. *crassa* (Brady), *Janischewskina minuscularia* Gan., *Tetrataxis* sp. (*T.* aff. *lata* Bogush et Juferev), *Palaeotextularia* sp., and *Eostaffella ovoidea* Rauser, all known from the upper Visean (many of them seem to be inherited from the underlying bed). Newcomers unknown below and appearing since the Serpukhovian Stage are *Neoarchaediscus postrugosus* (Reitl.), *N. akchimensis* (Grozd. et Leb.), and *"Millerella" tortula* Zeller. The last form occurs as single specimens. The interval is about 0.55 m thick.

Bed 4: detrital limestone similar to that from Bed 3 but more rich in microbioclastic fraction; ostracod remains are more abundant in this rock. Diversity of foraminifers is twice higher than before (35 species), and most abundant among them are species known from the upper Visean, e.g., *Endothyranopsis sphaerica* (Rauser et Reitl.), *Janischewskina minuscularia* Gan., *Parastaffella supressa* Schlyk., *Eostaffella tenebrosa* Viss., *Asteroarchaediscus ovoides* (Rauser), and species representing genera *Paraarchaediscus, Archaediscus*, and *Endostaffella.* Newcomers represent *Pseudoendothyra globosa* Ros., *Janischewskina delicata* Mal., (single specimens of both forms), *Planoendothyra* sp. (*P.* aff. *minima* Reitl.), and *Endothyra phrissa* (Zeller), all unknown below the Serpukhovian Stage. The bed is 0.8 m thick.

Bed 5: gray and hard, medium- to coarse-grained, bioclastic packstone; clasts are unsorted, representing mostly skeletal fragments and spines of brachiopods and echinoderms. Crinoid columnals are broken or powdered (indications of allochthonous burial).

Foraminifers represent 27 species, dominant among which are *Pseudoendothyra globosa* Ros., *"Millerella" tortula* Zeller, *Endothyranopsis sphaerica* (Rauser et Reitl.), and *Archaediscus krestovnikovi* Rauser, whereas other taxa occur as single specimens Especially remarkable is decrease in abundance of *Neoarchaediscus* forms in contrast to prevailing *Archaediscus* specimens. The bed is 0.6 m thick.

Fig. 2. Lithology, microfacies, and inferable sedimentation environments across the Visean–Serpukhovian boundary and during the Serpukhovian Age, the Zabor'e Quarry section: (1) limestone; (2) clayey limestone; (3) brachiopod limestone; (4) cherty limestone; (5) dolostone; (6) marl; (7) inaccessible for observation; (8) mudstone or clay.

Table 2. Distribution of foraminifers across the Visean–Serpukhovian boundary and throughout the Serpukhovian Stage, the Zabor'e Quarry locality

Stage	Visean				Serpukhovian																								
Horizon	Venev				Tarusa								Steshev														Protva		
Samples / Taxa	2a-1	2a-2	2a-3	2b	3a-1	3a-2	4	5	6	7	8	9	10	13	14	15	16	18	19, 20	21	26	28	29	48	49	51	53
Earlandia minor	+	+	+																								
Endothyranopsis sphaerica	+	+	+		+		+	+	+		+											+			+		
Endothyranopsis ex gr. *crassa*	+		+	+		+	+	+	+		+					x									+		
cf. *Omphalotis* sp.	+		+																								
Omphalotis exilis	+		+																								
Endostaffella parva	+	+		+		+	+															+					
Endostaffella shamordini	+						+	+			+	+										+					
"*Endostaffella*" *asymmetrica*	+	+		+	+	+	+	+	+	+	+					x	+	+				+					
Endostaffella fucoides	+	+		+																							
Howchinia bradyana	+																								+		
Eostaffella tenebrosa	+	+	+	+	?		+																				
Eostaffella ikensis	+																										
Eostaffella proikensis	+																										
Eostaffella ragushensis	+	+																							+		
Archaediscus krestovnikovi	+	+	+	?	?	+	+	+	+		+		+		+	+	+		+	+	+	+	+				
Archaediscus moelleri	+		+	+			+	+	+	+						+						+		+	+		
Archaediscus gigas	+																							+	+		
Paraarchaediscus embolicus	+																										
Asteroarchaediscus bashkiricus	+	+		+		+	+	+				+															
Palaeotextularia sp.	+		+			+		+	+													+			+		
**Neoarchaediscus* ex gr. *rugosus*	+	+		+		+	+	+	+		+	+	+	+	+	+	+	+		+	+	+			+		
Neoarchaediscus akchimensis	+			+		+	+		+		+					+						+					
Loeblichia paraammonoides		+												+		x											
Mikhailovella gracilis		+																									
Bradyina rotula		+	+																								
Globoendothyra globula			+	+																							
Omphalotis omphalotis			+																								
Janischewskina cf. *typica*			+	?																							
"*Eostaffella*" *parastruvei*			+				+			+	+											+		+			
Parastaffella propinqua			+																								
cf. *Climacammina* sp.			+	+																				+	?		
Janischewskina minuscularia				+		+	+									x											
Archaediscus nanus				+	+	+	+	+	+	+						x	+	+	+			+					
Eostaffellidae indet.					+																						
Earlandia vulgaris						+	+				+					x			+			+					
Pseudoammodiscus priscus						+	+	+			+	+	+			x				+							
Pseudoammodiscus volgensis						+	+				+					x								+			
Endothyra sp.						+	+	+	+	+	+					x	+	+					+	+	+		
"*Millerella*" *tortula*						+	+	+	+	+	+	+	+			x	+	+		+	?	+					
Neoarchaediscus postrugosus						+	+	+				+				x	+					+					
Eostaffella ovoidea						+					+	+										+					
Tetrataxis sp. (*T.* aff. *lata*)						+	+	+	+		+					x			+			+					
Archaediscus ex gr. *moelleri*						+	+	+	+		+		+			+		+				+		+	+		+
Earlandia elegans							+	+			+	+				x								+			
Pseudoendothyra globosa							+	+	+	+						+	?										
Planoendothyra aff. *minima*							+	+	+		+	+				x		+							+		

Table 2. (Contd.)

Stage	Visean	Serpukhovian																														
Horizon	Venev	Tarusa											Steshev																	Protva	
Samples / Taxa	2a-1	2a-2	2a-3	2b	3a-1	3a-2	4	5	6	7	8	9	10	13	14	15	16	18	19, 20	21	26	28	29	48	49	51	53
Parastaffella aff. *supressa*							+			+	+																
Janischewskina delicata							+				+														+		
Endothyranopsis intermedius							+	+	+													+			+		
Endothyra phrissa							+	+			+																
Asteroarchaediscus ovoides							+		+		+	+															
Paraarchaediscus stilus							+	+				+						+									
Paraarchaediscus koktjubensis							+	+																			
Endotaxis brazhnikovae							+	+	+		+	+				+		+				+		+			
Consobrinella consobrina							+	+	+		+					+		+				+					
Eostaffella spp.								+	+	+	+	+				x						+	+	+	+		+
"*Eostaffella*" *mirifica*								+	+		+					x	+	+						+	+		
Planoendothyra aljutovica									+		+					x						?			?		
Parastaffella sp. (sp. nov.)									+													+					
Pseudoendothyra sp. 1									+																		
Eogloboendothyra parva									+																		
Eostaffella ovesa											+	+				x											
Rectocornuspira sp.														+	+	x	?										
Betpakodiscus spp.														+	+	+	+	+	+		?	+			+		
Eostaffellina decurta																+	+	+				+		+	+		
Čepekia čepeki																x	?										
Loeblichia minima																+											
Loeblichia ukrainica																x											
Rectoendothyra latiformis																x	+					+					
Eostaffellina sp.																						+		+	+	+	?
Eostaffellina paraprotvae																						+		+	+	+	
Millerella pressa																						+					
Eostaffellina "*protvae*"																								+	+		
Eostaffellina subshpaerica																								+	+		
Eostaffellina shartimiensis																								+	+		
Eostaffella umbilicata																								+			
Deckerella sp.																								+			
Janischewskina adtarusia sp. nov.																									+		
Brenckleina rugosa																									+		
Planospirodiscus taimyricus																									+		
Pseudoendothyra parasphaerica																									?		

Symbols: (?) species defined with a doubt in oblique or incomplete sections; (X) species from additional sample collected 20 m away from marker of Bed 15 (see explanation in the text).

Remarks to taxa identification: (*) *Neoarchaediscus* ex gr. *rugosus* (Rauser) means a group of tightly associated species *N. rugosus* (Rauser) and *N. parvus* (Rauser) whose separate diagnosis is possible when ratio L/D (length to diameter of test) can be measured in strictly oriented axial sections, which are extremely rare (L/D ranges of 0.3–0.5 and 0.3–0.4 characterize the first and second form, respectively; Rauser-Chernousova, 1948d, 1948e). Taxa in quotes, such as "*Endostaffella asymmetrica* Rosovskaja, 1963,"*Eostaffella*" *mirifica* Brazhn., 1948, "*Eostaffella*" *parastruvei* (Rauser, 1948e), and "*Millerella*" *tortula* Zeller, 1953 (non *Zellerina* Mamet, 1970 in Mamet and Skip, 1970, p. 336; non *Zellerinella* Mamet, 1981) do not fit the generic diagnosis and should be revised.

Bed 6: organogenic-detrital limestone of the same composition and microfacies type; in comparison with Bed 5, abundance of bryozoan clasts increases, but echinoderm clasts are less frequent.

The diversity (26 species) and proportions between foraminiferal taxa change insignificantly, though *Earlandia, Pseudoammodiscus*, and *Paraarchaediscus* forms are missing among them. Two newcomers are *"Eostaffella" mirifica* Brazhn. and *Planoendothyra alyutovica* (Reitl.) Distinctly predominant in the assemblage are staffelliids with three-layer walls of tests. Among foraminifers inherited from the Visean Stage, there are single *Eogloboendothyra parva* (Tchern.). The bed is 1.3 m thick.

Bed 7: organogenic limestone with abundant smaller brachiopods; the rocks is classed with detrital packstone composed of coarse clasts of brachiopods, echinoderms, crinoids, ostracods, and gastropods. The diversity of foraminifers (8 species) considerably decreases (Table 2), and their tests occur as single specimens. The bed is 0.2 m thick.

Bed 8: organogenic limestone composed of algal remains (abundant dark microscopic inclusions of isometric shape, which are very hard). Under microscope, the rock was classed with algal-bioclastic packstone locally grading into microbioclastic wackestone. Thin-section areas of packstone type enclose rare ostracod shells and remains of bryozoans, echinoderms, brachiopods, and crinoids. Being concentrated and buried in the position of growth, remains of green algae Dasycladaceae form textural patterns of the boundstone type. Large rounded patches and dispersed particles of organic matter are clearly seen under microscope.

The assemblage of diverse foraminifers (26 species) is partially inherited from beds 4–6, but *Pseudoendothyra globosa* Ros. is already missing from this level (Table 2). Typical Serpukhovian species *Planoendothyra alyutovica* (Reitl.) and *"Millerella" tortula* Zeller (single specimens of both) continue their development and coexist with newly appearing *Eostaffella ovesa* Gan., *"Eostaffella" parastruvei* Rauser, *Pseudoendothyra* sp. 1, and Parastaffella sp. 1 (both taxa are rare). The bed is 1.2 m thick.

Bed 9: micro-grained aphanitic limestone classed under microscope with microbioclastic wackestone; large-size grains in the rock (1.5 to 2 mm across) are free of ooze (well washed), indicative of their allochthonous burial and transportation from the high-energy settings. Infrequent organic remains (less than 3 grains per one thin section) are those of echinoids, brachiopods, bryozoans, corals, thin-walled gastropod shells, and ostracods occurring in association with sponge spicules and algal bioclasts (single specimens).

The diversity of foraminifers (Table 2) decreases at this level (14 species). Their single tests are very small (0.14–0.20 mm). *Pseudoammodiscus* forms and *Neoarchaediscus* ex gr. *rugosus* (Rauser) are more frequent (2–3 specimens per one thin section) than other taxa. The bed is 11 m thick.

STESHEV HORIZON

Bed 10: homogeneous clayey aphanitic limestone of light gray color; texture of the rock is similar to that of limestone from Bed 9 except for the distinctive presence of microscopic laminae composed of fine terrigenous material. Foraminifers are twice less diverse (6 species) than below. *Pseudoendothyra globosa* Ros. has not been encountered here like in beds 8 and 9. The bed is 1.0 m thick.

Bed 13: pelitomorphic clayey limestone classed under microscope with microbioclastic wackestone; the rock encloses rare and large fragments of brachiopod shells. The assemblage of four foraminifers changes in composition: it includes new forms *Rectocornuspira* sp. and *Betpakodiscus* sp., and *Loeblichia paraammonoides* Brazhn. appears at this level once again. The bed is 0.8 m thick.

Bed 14: clayey limestone/compact limy marl classed under microscope with microbioclastic packstone; large grains in the rock (1–2 to 5 mm) are free of ooze (well washed and reworked) and suggest the allochthonous burial and transportation from the high-energy settings. The rock yields remains of echinoids, brachiopods, bryozoans, ostracods, and algae. Foraminifers (4 species) occur as single specimens (Table 2). The bed is sampled near its base and top, and both levels turned out to be identical. The bed is 0.4 m thick.

Bed 15: alternating clayey limestone and limy marl sampled at two levels; according to microscopic examination, microbioclastic packstone (Sample 15a-1) grades near the base into its variety bearing frequent clasts of medium to large size (2–3.5 mm), which are free of ooze, well washed, partially reworked, and oriented parallel to their long axes. Echinoderm and brachiopod remains are frequent. Other organic components represent single fragments of bryozoans, crinoid columnals, and sponge spicules. Foraminifers are rare, though three first tests of *Eostaffellina decurta* (Rauser), one with distinct morphological features, have been detected at this level.

Near the top (Sample15a-2), microbioclastic packstone is rich in large skeletal fragments of echinoderms, brachiopods, and bryozoans, which associate with sponge spicules and rare algal remains. Samples for foraminiferal analysis have been collected in two points of the bed. *Eostaffellina decurta* (Rauser) is present in both assemblages, which are not identical in abundance and diversity of foraminifers.

At the point where the bed is marked, foraminifers are represented by 12 species only. Their assemblage is more diverse than below, and each of five thin sections studied reveals coexistence of *Pseudoendothyra globosa* Ros., *Eostaffellina decurta* (Rauser), *"Millerella" tortula* Zeller, and *Betpakodiscus* sp. The last form is

very frequent. Two newcomers are *Loeblichia minima* Brazhn. and *Čepekia čepeki* Vasicek et Ruzicka. unknown from underlying beds.

Twenty meters away from that point, foraminifers are more diverse (33 species) and abundant. All forms of Sample 15a-2, except for *Pseudoendothyra globosa* Ros., occur here as well and coexist with 21 other species (Table 2, symbol "x"). Most abundant in this case are *Loeblichia paraammonoides* Brazhn., *L. ukrainica* Brazhn., *L. minima* Brazhn., *Rectocornuspira* sp., and *Eostaffellina decurta* (Rauser). They associate with frequent *Endotaxis brazhnikovae* (Bogush et Juferev), *"Millerella" tortula* Zeller, and *Pseudoammodiscus priscus* (Rauser). The assemblage clearly experienced a renewal, because the late Visean forms are considerably less abundant and diverse here. The bed is 0.55 m thick.

Bed 16: clayey limestone with basal interlayers of limy fissile mudstone; this microbioclastic wackestone (according to microscopic examination) is rich in large clasts representing remains of brachiopods, crinoids, bryozoans, and ostracods. Foraminifers are less diverse than below (15 species), and *Archaediscus krestovnikovi* Rauser is their dominant form. Many species identified in Sample 15a-2 do not occur in this bed (Table 2). *Eostaffellina decurta* (Rauser), *"Millerella" tortula* Zeller, and *Rectoendothyra latiformis* Brazhn. occur as single specimens. The bed is 0.75 m thick.

Bed 18: limestone above basal layer of microlaminated mudstone (Bed 17) is similar to that of Bed 16; large bioclasts in this microbioclastic rock are broken and reworked. The diversity (13 species) and abundance of foraminifers decreases here. The bed is 0.25 m thick.

Beds 19 and 20: rock textures and microfacies compositions do not change in comparison with those observable in Bed 18. Rare bryozoan remains have been observed. Foraminifers are abundant in some thin sections, but their abundance rate is variable, and other preparations turned out to be lacking these fossils, which are not diverse (only 5 forms have been distinguished). Forms *Betpakodiscus* sp. are dominant among them, and sometimes (Bed 19, thin sections 4, 9, and 10; Bed 20, thin sections 5 and 7) their abundance rate corresponds to 10 specimens per a thin section. Beds are 0.55 and 0.75 m thick, respectively.

Bed 21: micro- to coarse-grained clayey limestone (of wackestone or packstone type under microscope) with very small thin-walled brachiopods; dominant in the coarse fraction are skeletal fragments of echinoids and brachiopods. Foraminifers (5 species) again occur as single specimens. They have been identified in three thin sections (nos. 3, 7, and 11) of 26 studied. The bed is 0.55 m thick.

Bed 22: dark gray microlaminated mudstone that has not been studied. The bed is 0.15 to 0.20 m thick.

Bed 23: microbioclastic wackestone with single foraminifers, among which poorly identifiable archaediscids are dominant. The bed is 0.15–0.20 m thick.

Bed 24: dark gray to black mudstone that has not been studied; the bed is 0.1 m thick.

Bed 25: clayey limestone (microbioclastic wackestone under microscope) with frequent vertical fucoids and large bioclasts (0.4–0.5 mm). Foraminifers are represented by rare Archaediscaceae having tests with thin hyaline walls. The bed is 0.7 m thick.

Bed 26: clayey organogenic brachiopod limestone, a presumable analogue of the "*Lobatus* Mudstone Horizon" (Shvetsov, 1932). Under microscope, its detrital variety is classed with microbioclastic packstone. The foraminiferal assemblage (5 species) is similar to that of previous bed. The bed is 0.35 m thick. Rocks of beds 19–26 accumulated in sedimentation settings unfavorable for development of foraminifers (Fig. 2).

Bed 28: algal boundstone grading into bioclastic-algal packstone and displaying areas, where Dasycladaceae skeletons are buried in the position of growth. Microfacies are close to or repeat those of Bed 8. Foraminifers are abundant and diverse in the bioclastic-algal packstone in contrast to their rare occurrence in the algal boundstone. They represent nearly all 29 forms, which occur in underlying beds. First specimens of newly appearing *Eostaffellina paraprotvae* (Rauser) and *Millerella pressa* Thompson have been identified in four of 28 studied thin sections in association with *Eostaffellina decurta* (Rauser) whose frequent tests again appear at this level after a break. A poor preservation state of foraminiferal tests, many of which are broken and deformed suggest that their accumulation setting near the algal shoals were of a high-energy type. The bed is 0.20 m thick.

Bed 29: black microlaminated mudstone with thin laminae of clayey to marly limestone; the basal limy laminae yields abundant small thin-walled brachiopods (the "*Lobatus* Mudstone Horizon" after Shvetsov, 1932). Under microscope, the last rock type corresponds to brachiopod–crinoidal packstone or rudstone. Larger clasts representing remains of crinoids, brachiopods, and echinoids are closely packed. In its texture, the rock differs from all underlying varieties. Dominant among rare and uniform foraminifers are Archaediscaceae whose species *Archaediscus krestovnikovi* Rauser, *Endothyra* sp., and Eostaffella sp. have been identified. The bed is 2.1 m thick.

The upper part of Steshev Horizon (beds 30–47) is composed of compact mudstone intercalated with lenticular dolomite interlayers, which are lacking foraminifers. This interval is 8 m thick.

PROTVA HORIZON

Bed 48: light gray compact limestone (microbioclastic packstone or wackestone under microscope) with frequent or rare large bioclasts, which are well washed, free of ooze; identifiable in bioclasts are remains of corals, crinoids, bryozoans, echinoids, brachiopods, and ostracods.

Foraminifers from Bed 48 (20 species) represent a renewed assemblage that includes seven new forms, *Eostaffellina protvae* (Rauser), *E. subshpaerica* (Gan.), *Eostaffella umbilicata* Kir., *Deckerella* sp., *Climacammina* sp., and *Archaediscus gigas* Rauser inclided. *Eostaffellina decurta* (Rauser) and *E. paraprotvae* (Rauser) inherited from underlying beds of the Steshev Horizon are frequent at this level, but many species appearing in the upper Visean are absent. Single *Neoarchaediscus* forms are less diverse, while *Archaediscus gigas* Rauser mentioned above is frequent. Typical late Serpukhovian forms prevail over others, and *Eostaffellina* species dominate in microbioclastic wackestone. The bed is 0.15–0.20 m thick.

Bed 49: light gray compact microbioclastic wackestone and bioclastic packstone with large clasts; foraminifers frequent in the rocks represent 26 species, among which there are frequent late Visean species *Endothyranopsis sphaerica* (Rauser et Reitl.), *E. intermedia* (Rauser), and single *Howchinia bradyana* (Howchin). Frequent tests of *Janischewskina adtarusia* sp. nov. (msc.) with very thin walls (0.014 mm) appear once again. Two newcomers are *Brenckleina rugosa* (Brazhn.) and *Pseudoendothyra parasphaerica* Reitl., both unknown below the upper Serpukhovian Substage. The bed is 0.15 m thick.

Bed 51: microbioclastic wackestone with spines and clasts of brachiopod shells; among single foraminifers, there was identified *Eostaffellina* (?) sp. The bed is 1.2 m thick.

Bed 52: yellow plastic ferruginate clay with white patches of calcareous flour; the bed is 0.3 m thick.

Bed 53: thick-platy limestone of unclear texture; the rock encloses peloids and bioclasts of uncertain biological affinity. Rare foraminifers are represented by *Archaediscus* ex gr. *moelleri* Rauser, *Paraarchaediscus* sp., *Eostaffella* sp., and *Eostaffellina* (?) sp. whose identification is difficult. The bed is 0.3–0.4 m thick.

SEDIMENTATION SETTINGS

Comprehensive data on facies and paleogeography of the Moscow basin during the Serpukhovian time have been considered in a series of works by Osipova and Bel'skaya (1965, 1967, 1970, 1975) who described the used investigation methods and published extensive lists of references. They orally presented their considerations at the seminar on deep-drilling results in the Russian platform (Osipova and Bel'skaya, 1970, p. 107) and once informed that proper data are used in the Atlas of Lithological–Paleogeographic Maps of the USSR (Osipova and Bel'skaya 1967, p. 119).

These data based on "sectioning bed after bed" (Osipova and Bel'skaya, 1965, p. 6) contributed much to reconstruction of paleogeography of the Serpukhovian Age in the Moscow region. Unfortunately, they have been published in a generalized form and cannot be used in the case of Zabor'e section. Other investigations of Serpukhovian facies in the Moscow syneclise (Fomina, 1969; Makhlina and Zhulitova, 1984; Makhlina *et al.*, 1993) are inappropriate for the same reason. Nevertheless, by the visual examination of rocks in the Zabor'e Quarry, one can easily recognize the horizons and their boundaries in the section. Accordingly, let us consider below the resolution ability of the undertaken microfacies analysis for reconstruction of the carbonate sedimentation history.

When defining relationships between components of microfacies based on grain density in a standard thin section (2.5 × 2.5 cm), I met a difficulty, because the currently accepted models of microfacies and corresponding sedimentation environments operate with basins of a high biological productivity, e.g., with present-day basins of Florida, Triassic to Jurassic basins of Alps and Middle East, the Dinantian (Tournaisian–Visean) basin of Belgium, etc. (Ginsburg, 1953, 1956; Wilson, 1975; Fligel, 1982; *Sedimentary...*, 1990). At the same time, epicontinental basins of the Late Paleozoic have never been discussed widely.

Accordingly, I may infer in my case, with certain reservations, only three categories of sedimentation environments: (1) a shallow basin (persistent presence of foraminifers, algae, and remains of marine benthic organisms); (2) flooding events (abundance of microbioclasts in the ooze fraction and appearance in allochthonous burials of deeper-water faunal remains, e.g., of sponge spicule, juvenile ammonoids, and some benthic taxa; (3) shoaling events (distinct domination of algae remains, some of which are buried in the position of growth, absence of microbioclasts, and textures of grainstone and rudstone types). When interpreting energy factor during sedimentation, I took into account the density of microbioclasts in the ooze fraction and the preservation state of fossils, the frame-building taxa resistant to high-energy environments included.

Compositions and secular evolution of microfacies are described above. Generalizing them (Fig. 2), I infer the following evolution of sedimentation environments beginning since the terminal Visean till the Protva time.

Considerable changes in microfacies composition suggest that a distinct shoaling of the terminal Visean time gave place to transgression of the Serpukhovian Age.

The terminal Visean corresponded to a period of high- to medium-energy habitat environments of foraminifers. Bioclasts, pure ooze, and sparite of marine genesis represented main components of sediments.

In contrast, the low-energy sedimentation settings and predominance of microbioclsts among sedimentary components were characteristic of the initial Serpukhovian time. It is interesting therewith that the general taxonomic composition of faunal assemblages insignificantly changed across the Visean–Serpukhovian boundary, and prevailing below and above this level are remains of shelly and testaceous faunas, mainly of brachiopods and echinoids. Frame builders,

among them crinoids and bryozoans, were of subordinate significance in both periods. The corresponding section intervas are lacking corals and sponge spicules, which appeared in the mid-Serpukhovian time only. Algae *Calcifolium okense* Shwetz. et Bir. characteristic of the Visean gave their place to Dasycladaceae.

Alekseev *et al.*, (1996) recognized two successive episodes of shoaling in the epicontinental sea basin of the Serpukhovian time. The first one is recorded at the level delimiting the lower one third of the Tarusa Horizon, and the second happened in the mid-Steshev time without break in marine sedimentation.

Three transgressive phases separated by above episodes corresponded to the first half of the Tarusa time, to the terminal Tarusa and early Steshev periods, and to the initial Protva time.

Between the Tarusa–Steshev transgressive phase and Steshev shoaling, there is an interval of carbonate rock succession in the Zabor'e Quarry implying that sedimentation environments and energy parameters of foraminiferal habitats were most unstable during the corresponding period.

BIOSTRATIGRAPHIC UNITS

Foraminifers characterize a greater part of the Zabor'e section, being missing from an upper part of the Steshev Horizon only. Their impoverished assemblages are typical of the lower and upper thirds of that horizon as well (beds 13, 14, 19–26, and 29).

Variations in abundance and diversity of foraminifers reflect impacts of microfacies factor, sedimentation environments, and energy parameters of their habitat medium. The most diverse assemblages are confined to beds 3–6, 48, and 49 of medium-grained bioclastic wackestone and packstone, which accumulated in shallow-water calm settings, and to beds 2a, 8 and 28 of medium-grained grainstone and algal–bioclastic packstone deposited in a shallow zone with intermediate to high energy of medium.

The assemblages are impoverished in algal boundstone (beds 8, 28) and in sediments, which accumulated in periods of highly changeable and unstable environments (beds 2b, 3a, 19–27), and when the high influx of terrigenous material (beds 10–14, 19–27) was characteristic of sedimentation settings.

The increased diversity of cosmopolitan Serpukhovian forms in Bed 15, where they associate with abundant sponge spicules and juvenile ammonoids, reflects a moment, when the basin was deep and well connected with the open sea (Fig. 2).

Irrespective of microfacies changes, there are four levels, across which taxonomic renewals in foraminiferal assemblages were most significant. These levels separate five biostratigraphic units of foraminifers: *Eostaffella tenebrosa* Zone, *Neoarchaediscus postrugosus* Beds, and three successive *Pseudoendothyra globosa, Eostaffellina decurta*, and *Eostaffellina "protvae"* zones (Fig. 3).

Eostaffella tenebrosa Zone

Its lower boundary is undetectable. The zone yields foraminiferal species *Endothyranopsis sphaerica* (Rauser et Reitl.), *Janischewskina typical* Mikh., *Bradyina rotula* Eichw., *Howchinia bradyana* (Howchin), *Loeblichia paraammonoides* Brazhn., *Eostaffella tenebrosa* Viss., *Neoarchaediscus* ex gr. *rugosus* (Rauser), *Asteroarchaediscus bashkiricus* (Krest. et Theod.), and *Climacammina* sp., all typical of the uppermost Visean in the Moscow region. This unit corresponds to an upper part of the Venev Horizon.

Neoarchaediscus postrugosus Beds

The lower boundary is established at the level, where the diversity of typical late Visean foraminifers distinctly decreases. The assemblage consists of *Endothyranopsis sphaerica* (Rauser et Reitl.), *"Endostaffella" asymmetrica* Ros., *Archaediscus krestovnikovi* Rauser, *A. nanus* Rauser, and *Eostaffella tenebrosa* (?) Viss., which are inherited from the Venev Horizon and associate with newcomers *Neoarchaediscus postrugosus* (Reitl.), *N. akchimensis* (Grozd. et Leb.), and *"Millerella" tortula* Zeller frequent in the Tarusa Horizon of the studied section.

The unit is distinguished at the Tarusa Horizon base (Bed 3 only) and demonstrates a gradual transformation of foraminiferal assemblages by transition from the Venev to Tarusa Horizon, i.e., across the Visean–Serpukhovian boundary. I do not exclude however that *Pseudoendothyra globosa* Ros. appearing in Bed 4 may occur as well in Bed 3.

Pseudoendothyra globosa Zone

The lower unit boundary is placed at the appearance level of *Pseudoendothyra globosa* Ros. in Bed 4, and the upper one coincides with the first occurrence level of *Eostaffellina decurta* (Rauser) in Bed 15.

Three foraminifer taxa of the *Pseudoendothyra globosa* Zone, which appear at the commencement time of Serpukhovian Stage in the *N. postrugosus* Beds, associate with newly appearing forms. In addition to index species, the latter are represented by *Planoendothyra* sp. (*P.* aff. *minima*), *Janischewskina* cf. *delicata* Mal., *Endothyra phrissa* (Zeller), and *Endotaxis brazhnikovae* (Bogush. et Juferev). *"Eostaffella" mirifica* Brazhn. and *Planoendothyra aljutovica* (Reitl.) appear in this assemblage slightly higher, and all taxa associate with *Betpakodiscus* sp. near the upper unit boundary. Thus, the unit reveals a successive increase in diversity of foraminifers.

The newly appearing taxa of Serpukhovian Age coexist in the unit with typical late Visean forms known from underlying strata. Most frequent among the latter

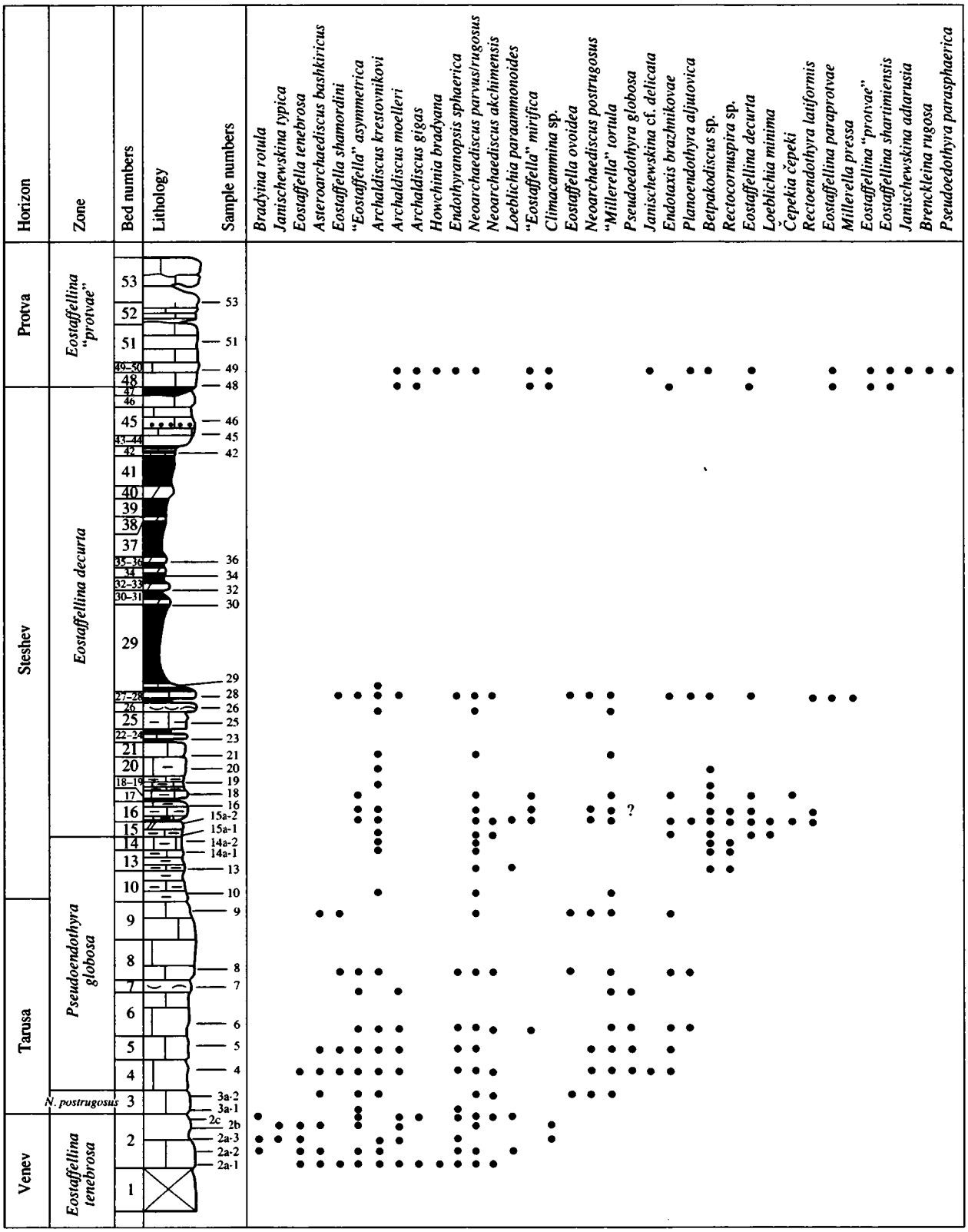


Fig. 3. Stratigraphic ranges of foraminiferal taxa and local biozones established in uppermost Visean–Serpukhovian interval of the Zabor'e Quarry section (symbols for lithology as in Fig. 2).

are *Endothyranopsis sphaerica* (Rauser et Reitl.), *"Endostaffella" asymmetrica* Ros., *Archaediscus* ex gr. *moelleri* Rauser, *A. krestovnikovi* Rauser, *A. nanus* Rauser, and *Neoarchaediscus* ex gr. *rugosus* (Rauser).

The assemblage is very diverse, with a gap in Bed 7, up to the Bed 8, where index species has not been encountered however. Above the last bed, diversity of foraminifers decreases down to four species (Table 2, Fig. 3). This appears to be a reaction of foraminifers to changes in sedimentation environments (Fig. 2).

The zone corresponds to the greater portion of the Tarusa Horizon (except for Bed 3) and also spans the lower interval of the Steshev Horizon (beds 4–14 of the latter). Thus, both boundaries of the unit do not coincide with the accepted delimitation levels of the indicated horizons.

Eostaffellina decurta Zone

The lower unit boundary marks the appearance level of its index species. First tests of the latter have been detected at the base of Bed 15 (Sample 15a-1).

Foraminiferal assemblage characteristic of the zone includes many forms present in the underlying unit (Table 2, Fig. 3). Taxa appearing here in addition to index species are *Rectoendothyra latiformis* Brazhn., *Loeblichia minima Brazhn.*, and *Čepekia čepeki* Vasicek et Ruzicka. Foraminifers are most diverse near the zone base. Their diversity significantly decreases in the middle portion, and the upper interval (Bed 28) demonstrates again the diversity increase. First appearance of *Eostaffellina paraprotvae* (Rauser) and *Millerella pressa* Thompson is recorded in the last interval that also demonstrates the recurrent occurrence of many taxa typical of the Serpukhovian Stage and upper Visean Substage, e.g., of *Endothyranopsis sphaerica* (Rauser et Reitl.), *Endothyranopsis crassa* (Brady), and many archaediscids.

The peculiar composition of the assemblage was controlled by unstable sedimentation environments and persistent changes in their habitat medium. The index species that has not been encountered in beds 19–26 and 29–46 continued its development in the next zone.

The described zone corresponds to the greater part of the Steshev Horizon (beds 15–47) in terms of stratigraphic subdivisions accepted in this work. Its upper boundary cannot be defined with a proper confidence, because foraminifers are either scarce in, or missing from beds 30–47 constituting the upper portion of the Steshev Horizon.

Eostaffellina "protvae" Zone

For the lower boundary of the unit, I accepted the first occurrence level of its index species in Bed 48. I do not exclude however that this form actually appears earlier and has not been encountered below because of unfavorable rock facies.

The zonal assemblage of high taxonomic diversity is considerably renewed, incorporating seven new foraminiferal taxa in addition to index species. Its characteristic late Serpukhovian forms are *Eostaffellina shartimiensis* (Mal.), *E. subshpaerica* (Gan.), *Brenckleina rugosa* (Brazhn.), *Deckerella* sp., *Pseudoendothyra parasphaerica* Reitl., *Eostaffella umbilicata* Kir., new species *Janischewskina adtarusia* Gibshman (msc.), and *Eostaffellina paraprotvae* (Rauser) that appears first below in Bed 28. Rare specimens of *Eostaffellina decurta* (Rauser) still occur at this level.

Taxa typical of the upper Visean and once again appearing in the unit are *Endothyranopsis* ex gr. *crassa* (Brady), *E. sphaerica* (Rauser et Reitl.), *E. intermedia* (Rauser), *Howchinia bradyana* (Howchin), and *"Eostaffella" parastruvei* Rauser. Representatives of the genus *Eostaffella* are abundant (*E. ovoidea* Rauser). Tests of *Archaediscus moelleri* Rauser. prevail among archaediscids. Another characteristic form is *Climacammina* sp., but *"Millerella" tortula* Zeller, *Neoarchaediscus postrugosus* (Reitl.), and *N. akchimensis* (Grozd. et Leb.) have not been encountered. *N.* ex gr. *rugosus* (Rauser) and *Betpakodiscus* forms occur as single specimens.

The zone corresponds to the Protva Horizon. Its assemblage looks rather mature and demonstrates the diversity increase of the genus *Eostaffellina* from Bed 48 to Bed 49. Similar assemblages are known from other sections of the Moscow syneclise (Fomina, 1977). Species *Brenckleina rugosa* (Brazhn.) appearing in Bed 49 and unknown in the Donetsk basin below the upper portion of the Protva Horizon is typical of the Zapaltyube Horizon (Aizenverg *et al.*, 1983). It implies that this portion of the Protva Horizon is exposed in the Zabor'e Quarry. Besides the Donetsk basin, *Brenckleina rugosa* (Brazhn.) is known from the uppermost Chesterian sequence of North America and occurs in sections of southern Illinois from the base of Menard Limestone up to the boundary with the Grove Church Shale (Brenckle, 1990).

PROBLEMS OF FORAMINIFERAL ZONATION IN THE SERPUKHOVIAN STAGE

Biozonation of foraminifers that has been distinguished in the Zabor'e Quarry corresponds, though with modifications, to their original biozonation suggested for the Serpukhovian Stage by Reitlinger (see in Lipina and Reitlinger, 1970) and discussed in some later works (Osipova and Bel'skaya 1975, p. 11). It was doubted afterward (Vdovenko *et al.*, 1989, 1990) and excluded from the general and regional stratigraphic schemes of the Serpukhovian Stage accepted for the Russian platform (*Postanovleniya ...*, 1982). It is necessary to mention here that all versions suggested after the original one are based on generalized data and lack references to the concrete sections. Accordingly, it is impossible to evaluate the corresponding criteria used to establish boundaries between foraminiferal zones

Table 3. Evolution of viewpoints on foraminiferal zonation in the Serpukhovian Stage (×)

Stage	Horizon	Foraminiferal zones of general and regional* schemes for the Russian platform			Foraminiferal zones in the Zabor'e section	
		Lipina and Reitlinger, 1970	*Postanovleniya...*, 1982; Kagarmanov, 1998	Vdovenko *et al.*, 1990; *Reshenie...*, 1990**	Makhlina *et al.*, 1993	This work
Serpukhovian	Zapaltyube	*Eosigmoilina, Monotaxinoides, Eolasiodiscus***	*Eostaffellina protvae – Eosigmoilina explicata – Monotaxinoedes subplanus*	*Eosigmoilina explicata – Monotaxiniodes subplanus*	Rocks are missing	Rocks are missing
	Protva	*Eostaffella protvae*		*Eostaffellina protvae*	*Eostaffellina protvae*	*Eostaffellina "protvae"*
	Steshev	*Eostaffella decurta*	*Pseudoendothyra globosa – Neoarchaediscus parvus*	*Pseudoendothyra globosa – Neoarchaediscus parvus*	*P. globosa – N. parvus*	*Eostaffellina decurta*
	Tarusa	*Pseudoendothyra illustria-P. globosa*				*P. globosa*
						*N. postrugosus***
Visean	Venev	*Eostaffella tenebrosa*	*Endothyranopsis Òrassa – Archaediscus gigas* (upper part)	*Eostaffella tenebrosa – Endothyranopsis sphaerica*	Unstudied	*Eostaffella tenebrosa*

(×) Serpukhovian Stage range is after Kagarmanov, 1998; * regional scheme for the Russian platform; ** beds with foraminifers.

and their correlation with boundaries between horizons. In contrast, the original zones (Lipina and Reitlinger, 1970) are easily recognizable in well-documented sections.

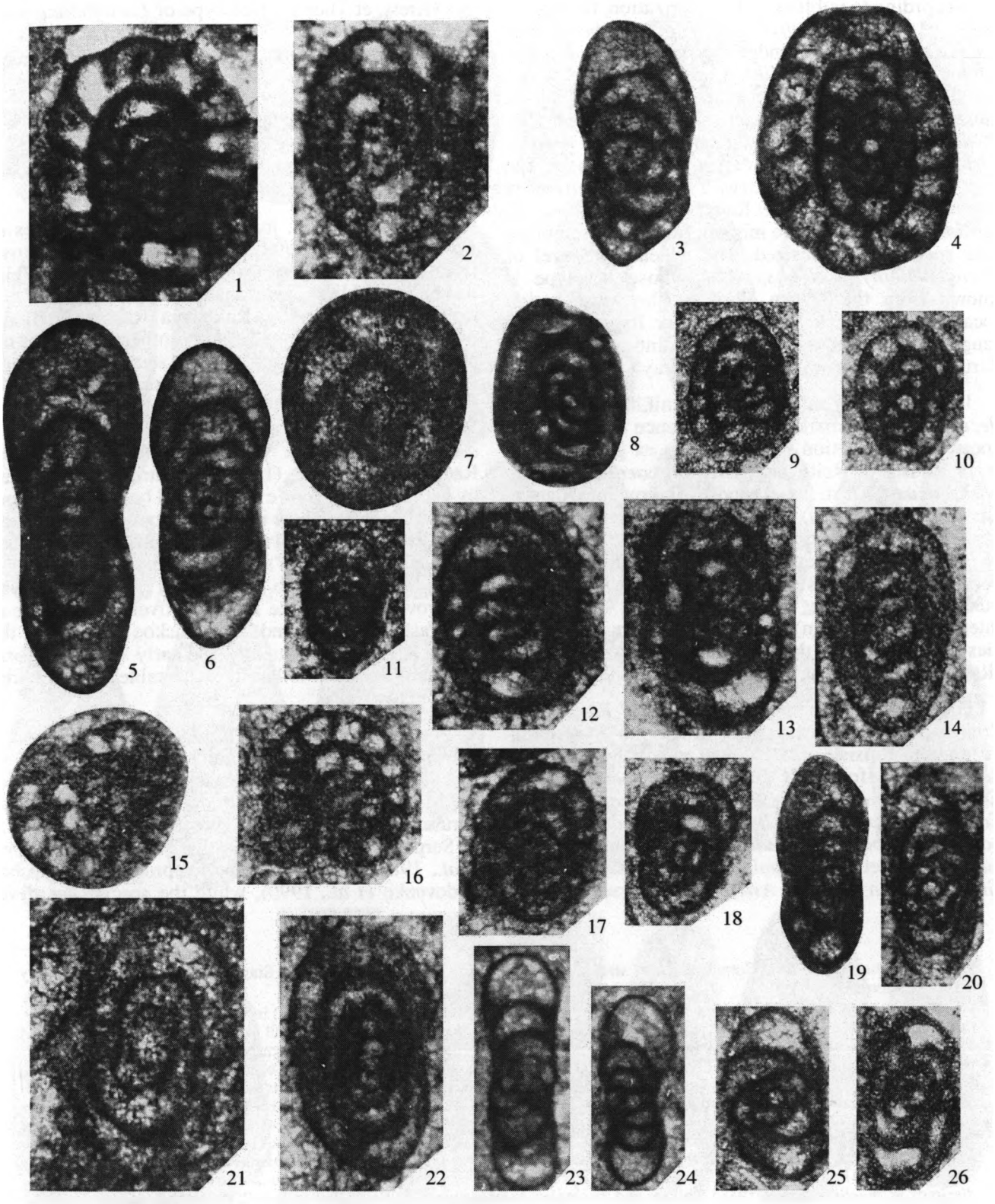
Reitlinger who was first to subdivide the Serpukhovian Stage of the Russian platform into foraminiferal zones (Lipina and Reitlinger, 1970) correlated them with coeval units of North America and the Donetsk and Franco-Belgian basins (Table 3). The *Pseudoendothyra illustria–P. globosa* (Tarusa Horizon), *Eostaffella decurta* (Steshev Horizon), and *Eostaffella protvae* (Protva Horizon) zones distinguished in the stage have been regarded as correlative with the *Eumorphoceras* Genus Zone of Franco-Belgian basin (Conil *et al.*, 1963) and with zones CI Na–Cl Nb of the Donetsk basin (Aizenverg *et al.*, 1963) or with an upper part of the upper Chesterian *Paramillerella* Zone of North America (Zeller, 1953).

Biozonation under consideration was based on the original data of Reitlinger and on results obtained in collaboration with Rauser-Chernousova (1936, 1940, 1948). Data published by Rozovskaya (1963) had been taken into account as well.

All zones, which differed from each other in their fossil content, had been regarded as equal in ranges to corresponding horizons. Unfortunately, neither type sections, nor boundary stratotypes had been indicated.

Plate I. Fusulinidae of stratigraphic significance from the Serpukhovian Stage, the Zabor'e Quarry locality (magnification 150 except for indicated otherwise):

(1, 2) *Pseudoendothyra globosa* Rosovskaja, 1963, oblique sections of embryonic chambers, Bed 6, thin sections 16 and 23. (3, 4, 6, 19, 20) *"Millerella" tortula* Zeller, 1953: (4) complete section normal to the coiling axis, ×170; (6) complete axial section, ×170; (3) ×170 and (20) oblique sections crossing the embryonic chamber; (19) ×170, oblique section crossing chambers 4 and 11 of the last whorl and embryonic chamber (see complete section); specimens from Bed 3a-2, thin sections 2k and 2 (4, 19) and from Bed 4, thin section 13 (20). (5) *Millerella pressa* Thompson, 1944, complete axial section, ×170, Bed 28, thin section 4". (7–11) *Eostaffellina decurta* (Rauser, 1948): (7) a fragment of last whorl, three-layer test wall with tectum and upper and lower tectoria, ×300; (8) ×170 and (11) complete axial sections; (9,10) tangential sections; (8) Bed 15 1, thin section 5; (7, 9–11) Bed 15 -2, thin sections 4, 5, 9, 11. (12–14, 22) *Eostaffellina paraprotvae* (Rauser, 1948d): 22 - complete axial section, ×170, Bed 49, thin section 4; (12–13) tangential and (14) slightly oblique axial sections, Bed 48, thin sections 5, 2, 13. (15–17) *Eostaffellina "protvae"* (Rauser, 1948): (15, 16) incomplete transverse and (17) slightly oblique axial sections, all specimens from Bed 48, thin sections 5, 7. (18) *Eostaffellina protvae–E. decurta*, Bed 48, thin section 6. (21) *Eostaffellina shartimiensis* (Malakhova, 1956), axial section, Bed 48, thin section 7. (23, 24) *Endostaffella shamordini* (Rauser, 1948): (23) axial and (24) high axial sections; both specimens from Bed 4, thin section 3. (25, 26) *"Endostaffella" asymmetrica* Rosovskaja, 1963, axial sections: (25) from Bed 3, specimen 3a-1, thin section 4; (26) from Bed 6, thin section 10.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According to published characterization, the assemblage of *Pseudoendothyra illustria–P. globosa* Zone includes, in addition to index species, *Endothyranopsis sphaerica* (Rauser et Reitl.), *Endothyra bradyi* Mikh., *E. kirgisana* Rauser, *Endostaffella parva* (Moeller), *Eostaffella parastruvei* Rauser, *E. pseudostruvei* Kir., *Neoarchaediscus parvus* (Rauser), *N. rugosus* (Rauser), and *Archaediscus krestovnikovi* Rauser. The fact that *Globivalvulina parva* Tschern., *Bradyina* ex gr. *cribrostomata* Rauser et Reitl., and *Eostaffellina* aff. *paraprotvae* (Rauser) are missing from the assemblage was specially emphasized. The appearance level of *Pseudoendothyra globosa* Ros. whose holotype is known from the Tarusa Horizon (the Meshcherino locality) was indicated as zonal marker. Its stratigraphic range was regarded as equal to the joint interval of the Tarusa and Protva horizons (Rozovskaya, 1963).

In distinction from the previous unit, the *Eostaffella decurta* Zone is marked by appearance of this index species in association with *Bradyina* ex gr. *cribrostomata* Rauser et Reitl. and *Asteroarchaediscus* ex gr. *bashkiricus* (Krest. et Theod.). Following Rauser-Chernousova (1948a), Reitlinger (1963) regarded *E. decurta* (Rauser) as the ancestral form of *Eostaffellina* ex gr. *protvae* (Rauser). Holotype of *Eostaffellina decurta* (Rauser) is known from the lower Serpukhovian Substage (borehole Syzran' 401, depth interval of 771–780 m). Stratigraphic range of this species corresponds to the Steshev and Protva horizons (Rauser-Chernousova, 1948a).

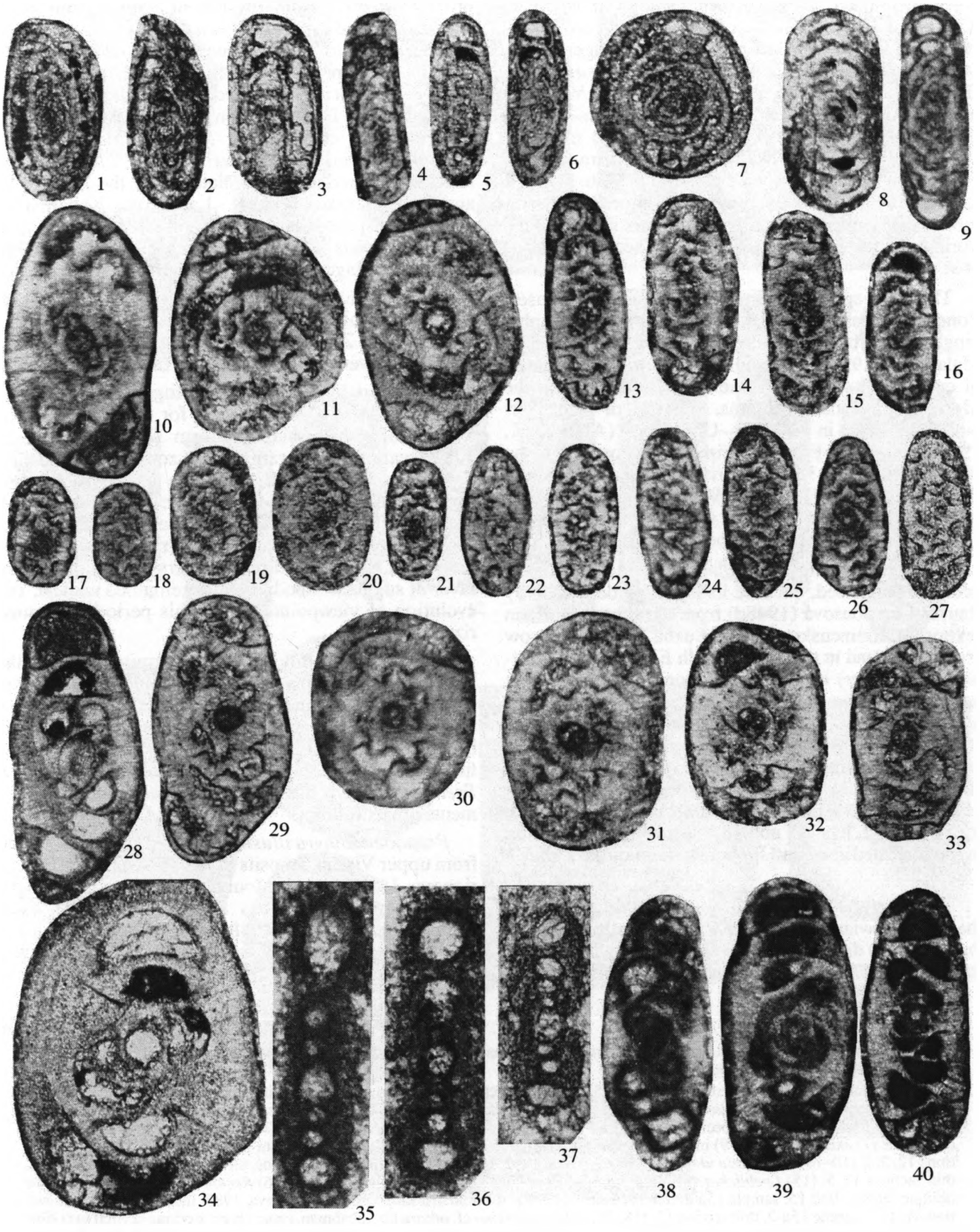
Foraminiferal assemblage of *Eostaffella protvae* Zone is more diverse than those of two previous subdivisions and consists of late Visean and newly appearing Serpukhovian forms. It characterizes coexistence of *Eostaffella parastruvei* Rauser, *E. ikensis* Viss., *Eostaffellina protvae* (Rauser), *Pseudoendothyra globosa* Ros., *P. parasphaerica* Reitl., *Bradyina cribrostomata* Rauser et Reitl., *Janischewskina calceus* (Gan.) *Globivalvulina* sp., and *Asteroarchaediscus bashkiricus* (Krest. et Theod.). Holotype of *Eostaffellina protvae* (Rauser) is known from the lower part of the Protva Horizon (the Kremenskoe locality). Distribution range of this species corresponds to the Protva Horizon (Rauser-Chernousova, 1948). As is shown above, all three zones are well distinguishable in the Zabor'e Quarry, although they have been established without reference to concrete sections, in which the corresponding biozonation can be observed.

The only section, in which distribution ranges of *Eostaffellina decurta* (Rauser), *Eostaffellina protvae* (Rauser), and *Bradyina cribrostomata* Rauser et Reitl. had been established, was that of borehole Syzran' 401 drilled in the Samarskaya Luka area (Rauser-Chernousova *et al.*, 1940). Limited data on other sections of the Moscow basin elucidated only a set of foraminiferal species distributed across the Visean–Serpukhovian boundary, distribution patterns of Archaediscidae in the Serpukhovian Stage, and appearance level of *Pseudoendothyra illustria* Viss. According to remark of Rauser-Chernousova (1948a), foraminifers collected by her from the Zabor'e Quarry had been studied more comprehensively than in other sections sampled by Shvetsov and Sarycheva (Venev Monastery site at the Osetr River, Sukhodol'sk Quarry near Aleksin, Tarusa River sites, Bekhovo Quarry at the Oka River, Toporovo Ravine at the Besputa River, Zaikino site at the Vashana River, and Kremenskoe locality at the Luzha River). Consequently, the early version of foraminiferal zonation is hardly acceptable for the entire Russian platform.

For instance, *Pseudoendothyra globosa* Ros. is known, besides the type locality, only from the Tarusa and Steshev horizons of northern calm-water zone of the Moscow syneclise (Fomina, 1969) and from the Tarusa Horizon of the Novogurovskii reference section of Serpukhovian deposits in the Tula area (Makhlina *et al.*, 1984). Its holotype was reproduced only once (Vdovenko *et al.*, 1990), while the species has never

Plate II. Archaediscidae and Cornuspiridae of stratigraphic importance from the Serpukhovian Stage, the Zabor'e Quarry locality (magnification150).

(1–3) *Brenckleina rugosa* (Brazhnikova, 1964): (1, 2) axial sections, Bed 49, thin section 8; (3) incomplete axial section, Bed 49, thin section 1. (4–7) *Betpakodiscus* sp.: (4–6) axial sections, (7) section normal to coiling axis; all forms from Bed 49, thin sections 1, 2, and 11. (8, 9) *Planospirodiscus taimyricus* Sosipatrova, 1962: (8) tangential section, (9) nearly axial section; both specimens from Bed 49, thin sections 3 and 5. (10, 31) *Neoarchaediscus* aff. *akchimensis* (Grozdilova et Lebedeva, 1954): (10) axial section, Bed 4, thin section 20; (31) oblique section, Bed 15a-2, thin section 5. (11, 12) *Neoarchaediscus akchimensis* (Grozdilova et Lebedeva, 1954), axial sections: (11) Bed 3, sample 3a-2, thin section 4; (12) Bed 4, thin section 9. (13–16) *Neoarchaediscus postrugosus* (Reitlinger, 1949): (13) axial section, Bed 3, sample 3a-2, thin section 9; (14–16) nearly axial sections; (14, 15) Bed 3, sample 3a-2, thin section 1; (16) Bed 6, thin section 6. (17–20) *Asteroarchaediscus bashkiricus* (Krestovnikov et Theodorovich, 1947), axial sections, all from Bed 3, sample 3a-2, thin sections 4 and 6. (21–27) *Neoarchaediscus* ex gr. *rugosus* (Rauser, 1948d, 1948e), nearly axial sections: (21–26) Bed 3, sample 3a-2, thin sections 11, 10, 6, 9, 11; (27) Bed 15, sample 15a-2, thin section 2. (28, 29) *Archaediscus* ex gr. *moelleri* Rauser, 1948d, oblique and incomplete axial sections (28) Bed 6, thin section 2; (29) Bed 3, sample 3a-2, thin section 10. (30) *Asteroarchaediscus ovoides(?)* (Rauser, 1948d), nearly axial section, Bed 4, thin section 5. (32, 33) *Archaediscus nanus* Rauser, 1948d, nearly axial complete sections: (32) Bed 18, thin section 2; (33) Bed 3, sample 3a-1, thin section 4. (34) *Archaediscus gigas* Rauser, 1948d: incomplete axial section, Bed 49, thin section 9. (35, 36) *Rectocornuspira (?)* sp. (Wartin, 1930 in original description, wall structure is unknown), nearly axial section in spiral part and oblique section in straightened part, both from Bed 15a-2, thin sections 5, 9. (37) *Čepekia čepeki* Vasicek et Ruzicka, 1957, axial section, Bed 15a-2, thin section 1. (38) *Paraarchaediscus stilus* (Grozdilova et Lebedeva, 1956): oblique axial section, Bed 4, thin section 13. (39, 40) *Archaediscus krestovnikovi* Rauser, 1948e; two oblique axial sections, Bed 6, thin sections 6, 8.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been mentioned in other publications devoted to description of Serpukhovian foraminifers from southern areas of the Russian platform (Reitlinger *et al.*, 1977; Fomina, 1977), the Urals (Sultanaev *et al.*, 1978; Grozdilova *et al.*, 1975; Postoyalko, 1990), Voronezh massif (Aizenverg *et al.*, 1968), and Timan–Pechora province (Durkina, 1990). New specimens of *Pseudoendothyra globosa* Ros. that is known from sections of Moscow syneclise only are pictured in Plate I (figs. 1 and 2), although Vdovenko who reexamined foraminifers of the Serpukhovian Stage from this structure and Voronezh anteclise failed to identify it (see in Makhlina *et al.*, 1993).

The index species of *Eostaffellina decurta* (Rauser) Zone and Steshev Horizon is of a wider geographic range. It has been described for the second time by Malakhova (1956) and represents a form characteristic of calm-water deposits of the Steshev Horizon in the Moscow syneclise (Fomina, 1969), of the Serpukhovian Stage in the Volga–Ural region (Aliev *et al.*, 1975), and of the Protva Horizon in exposures of the Bol'shoi Kizil and Khudolaz rivers in eastern areas of the southern Urals (Ivanova, 1973). Other foraminifers typical of the Zabor'e Quarry section (Plate I, figs. 7–11) have never been pictured before.

The index species *Eostaffellina protvae* (Rauser) needs to be revised, because its holotype described by Rauser-Chernousova (1948d) from the Protva Horizon (exposure Kremenskoe at the Luzha River, Moscow region) is found in association with *Eostaffellina paraprotvae* (Rauser) in one thin section, and represents a tangential section of the latter, as I established revising her collection (GIN no. 2853). This taxon in understanding of Reitlinger (1963, Plate 1, fig. 2) has been regarded as a marker of the Protva Horizon and its stratigraphic analogues (Marfenkova, 1991; Rumyantseva, 1989, Kulagina *et al.*, 1992), and exactly these forms (Plate I, figs. 17 and 18) have been encountered in the indicated zone and Protva Horizon of the Zabor'e Quarry.

The currently authorized foraminiferal zonation of the Serpukhovian Stage in the Russian platform (*Reshenie...*, 1990) differs from that suggested by Lipina and Reitlinger (1970). For the Tarusa and Steshev horizons problematically distinguishable in other regions, there was authorized a single *Pseudoendothyra globosa–Neoarchaediscus parvus* Zone, whereas the *Eostaffellina protvae* Zone corresponding to the Protva Horizon was left without changes. According to correlation with sections of the Donetsk basin, the Serpukhovian Stage zonation was crowned by *Eosigmoilina explicata–Monotaxinoides subplanus* and *Plectostaffella bogdanovkensis* zones correlated with the Zapaltyube and Voznesensk horizons. Later on, Kagarmanov (1998) transposed the last horizon and *Plectostaffella bogdanovkensis* Zone from the Serpukhovian into Bashkirian Stage (Table 3).

The interval corresponding to the Zapaltyube Horizon and *E. explicata– M. subplanus* Zone is likely missing from the Zabor'e section being eliminated during an erosion event of the Bashkirian time.

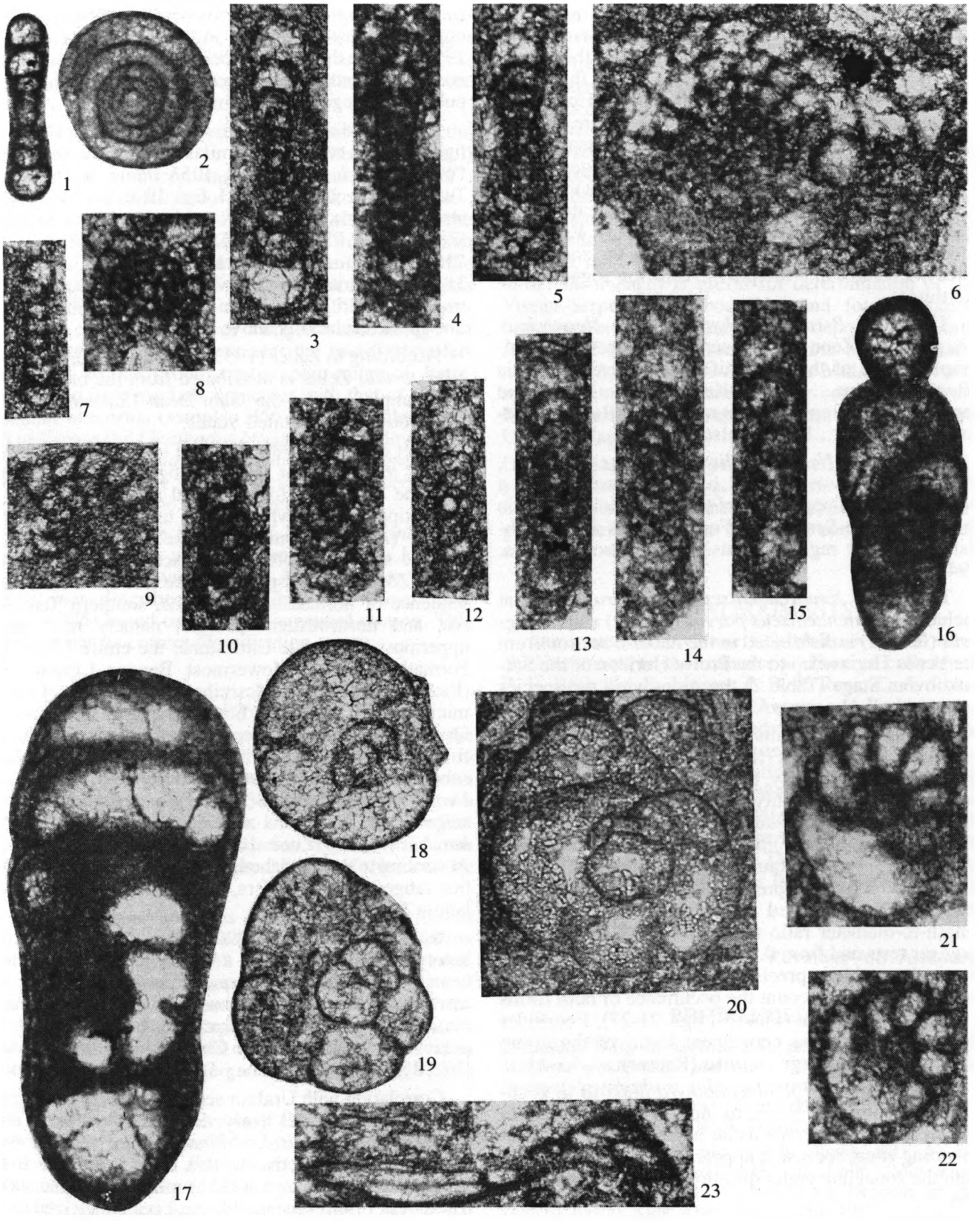
After work by Lipina and Reitlinger (1970) and later authorization of a general scale for the Carboniferous System of the Russian platform (*Postanovleniya...*, 1982), variants of foraminiferal zonation in the Serpukhovian Stage were discussed in a series of publications (Vdovenko *et al.*, 1989, 1990; *Reshenie...*, 1990; Makhlina *et al.*, 1993), but nobody, except for Makhlina and her colleagues, referred to concrete sections with well-studied foraminifers, and arguments in favor of suggested modifications remained unclear. The evolution of viewpoints during this period is summarized in Table 3.

The results of this work on foraminifers from the Zabor'e Quarry imply that the early variant of zonation (Lipina and Reitlinger, 1970) is quite realistic despite the dissimilar levels of boundaries distinguishable between zones and horizons. I think however that the first index taxon should be excluded from the name *Pseudoendothyra illustria–P. globosa* Zone. The arguments are as follows.

Pseudoendothyra illustria Viss. originally described from upper Visean deposits of the Tuimaza area (Vissarionova, 1948) was also found in the Mikhailovo Horizon of borehole sections Syzran' 401, 10 and 22 (Rauser-Chernousova, 1948c) and listed among foraminifers typical of the Protva Horizon in summary tables characterizing distribution of these fossils in the

Plate III. Some foraminifers of stratigraphic importance from the Serpukhovian Stage, the Zabor'e Quarry locality (magnification 150 except for otherwise indicated).

(1, 2) *Pseudoammodiscus volgensis* Reitlinger, 1951: (1) axial section; (2) section normal to coiling axis; both from Bed 15, sample 15a-2, thin section 4. (3–6) *Loeblichia paraammonoides* Brazhnikova, 1962: (3, 4) axial sections, (5) oblique axial section, (6) section normal to coiling axis; all specimens from Bed 15, sample 15a-2, thin sections 2, 7, 4, 12. (7–9) *Loeblichia minima* Brazhnikova, 1962: (7) axial section, (8, 9) incomplete sections normal to coiling axis; all specimens from Bed 15, sample 15a-2, thin sections 12, 2, 2. (10–14) *Loeblichia ukrainica* Brazhnikova, 1962, incomplete oblique axial sections, all from Bed 15, sample 15a-2, thin sections 12, 5. (15) *Loeblichia?* sp., high section, Bed 15, sample 15a-2, thin section 5. (16) *Rectoendothyra* sp., incomplete oblique section, Bed 15, sample 15a-2, thin section 5. (17) *Rectoendothyra latiformis* Brazhnikova, 1983, incomplete oblique section, Bed 15, sample 15a-2, thin section 12. (18, 20) *Janischewskina* cf. *adtarusia* Gibshman, (msc.) high sections normal to coiling axis, ×36, both from Bed 49, thin sections 12, 2. (19) *Janischewskina* cf. *delicata* Malakhova, 1956, section showing four chambers of the first whorl and two chambers of the second one, ×60, Bed 4, thin section 21. (21, 22) *Endotaxis brazhnikovae* (Bogush et Juferev, 1962), sections normal to coiling axis: (21) Bed 15, sample 15a-2, thin section 12; (22) Bed 5, thin section 15. (23) *Tetrataxis* sp., exotic form, Bed 15, sample 15a-2, thin section 14.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southern limbs of Moscow syneclise and in the central Russian platform and Cis-Urals (Rauser-Chernousova, 1948a, 1948b). Thus, stratigraphic range of this poorly identifiable species spans interval from the upper Visean Mikhailovo Horizon to the Protva Horizon of the Serpukhovian Stage inclusive. The picture of a single, originally described specimen (Vissarionova, 1948) was reproduced later on by Reitlinger (1963, Plate 2, fig. 9). Accordingly, this form of a wide stratigraphic range, greater than that of the indicated zone, needs to be studied more comprehensively. Exclusion of *Pseudoendothyra illustria* Viss. from the zone name does not change in principle the zonation suggested by Reitlinger.

The name *Pseudoendothyra globosa–Neoarchaediscus parvus* Zone that is accepted in the general stratigraphic scale of the Carboniferous System in Russia (*Postanovleniya...*, 1982; Kagarmanov, 1998) and repeated in the regional scheme for the Russian platform (*Reshenie...*, 1990) is also lame.

Describing *Neoarchaediscus parvus* (Rauser), Rauser-Chernousova (1948d, p. 233) pointed out that it occurs from the upper part of the Mikhailovo Horizon throughout the Serpukhov Formation, as it is established in other regions (Rauser-Chernousova, 1948a, 1948b).

The group *Neoarchaediscus* ex gr. *rugosus* that includes *Neoarchaediscus parvus* (Rauser) and *N. rugosus* (Rauser) is distributed in the Zabor'e section from the Venev Horizon up to the Protva Horizon of the Serpukhovian Stage (Table 2) thus approving the species appearance in the upper Visean Substage.

In addition, the results of revision of pictured holotype and two paratypes (Brenckle et al. 1993, Plate 16, fig. 10, Aleksin locality, Serpukhovian Stage; figs. 11 and 12, Bekhovo locality, Venev Horizon) imply that diagnosis of *Neoarchaediscus parvus* (Rauser) and limits of its variability do not differ from those characterizing the *N. rugosus* (Rauser) type species, and that the former taxon likely represents a lesser morphotype of the latter. As indicated in the first description, the length-to-diameter ratio ranges from 0.3 to 0.5 in *N. rugosus* tests and from 0.3 to 0.4 in those of *N. parvus* that is detectable in precisely oriented axial sections of tests. Taking into account the occurrence of both forms in a tight association (Plate II, figs. 21-27), I consider them in this work as constituent forms of the group *Neoarchaediscus* ex gr. *rugosus* (Rauser).

Thus similar to *Pseudoendothyra illustria* in zonation by Reitlinger, it seems unreasonable to regard *Neoarchaediscus parvus* as an index species of corresponding zone, because it appears below basal levels of both the zonal unit under question and the Tarusa Horizon.

At the end of current discussion, I would like to mention that Reitlinger correlated three foraminiferal zones of the Serpukhovian Stage in the Russian platform with the *Paramillerella* Zone (*Millerella, Zellerina*, or *Zellerinella* in various interpretation) of the upper Chesterian sequence in North America (Zeller, 1953). Before the present work, American taxa have not been identified among foraminifers from the Serpukhovian Stage of Russia.

In the studied collection, there are forms (Plate I, figs. 3, 4, 6), which are similar and even identical to *"Millerella" tortula* Zeller, 1953 (Plate 26, fig. 25). Two other specimens (Plate I, figs. 19 an 20) likely represent the same species pictured in the oblique section (Zeller, 1953, Plate 26, fig. 18). In the Zabor'e section, *"Millerella" tortula* Zeller appears in Bed 3 (Sample 3a-2) and marks, together with *Neoarchaediscus postrugosus* (Reitl.), the Visean–Serpukhovian boundary. The species is lacking above Bed 28.

According to Zeller (1953, p. 190, Fig. 5), *"Millerella" tortula* Zeller is distributed from the base to the top (sample 1–4) of the Glen Dean Limestone in its type section of the United States.

Later on, Rich (1980) pictured specimens of *"Millerella" tortula* Zeller (Plate14, figs. 23, 24, 27, 28) from the Zone 17 in foraminiferal zonation by Mamet and Skipp (1970), which were termed as *Zellerina tortula* (Zeller) resembling some species from the Zabor'e section according to description. As is indicated, Zone 17 corresponds to the middle Chesterian sequence of northeastern Alabama, southern Tennessee, and northwestern Georgia, namely to "...the uppermost Monteagle Limestone, the entire Hartselle Formation, and the lowermost Bangor Limestone" (Rich, 1980, p. 8). The described assemblages of foraminifers from the Serpukhovian Stage of the Zabor'e Quarry include many species in common with those from Zones 17 and 18 of North American stratigraphic subdivisions mentioned above (Rich, 1980). These are *Neoarchaediscus parvus* (Rauser) (both zones), *N. postrugosus* (Reitl.) (both zones), *Planoendothyra alyutovica* (Reitl.) (Zone 18), and some other forms. According to the published table summarizing distribution ranges of foraminifers, these species do not occur below Zone 17.

Recently Brenckle (1990) reproduced images of holotype and paratypes of *"Millerella" tortula* Zeller from the type Chesterian sequence and defined them as markers of Horizon 6 at the base of the Glen Dean Formation. Thus, the direct correlation between the Serpukhovian Stage and upper Chesterian subdivisions of United States appears to be possible.

Correlation with Uralian sections is more problematic, because typical early Serpukhovian taxa are unknown there. *Janischewskina delicata* Mal. is the only species perspective in this respect. By the first description, Malakhova (1956) pointed to the late Visean age of this species that has been authorized as a Serpukhovian form by reexamination (Reitlinger in *Stratigrafiya...*, 1973). Revising thin sections with holotype and paratype of *J. delicata* Mal., I identified several Serpukhovian species *Biseriella parva* Tschern.

missed by Malakhova (1956) from the assemblage composition. Accordingly, the conclusion of Reitlinger that the first occurrence level of *J. delicata* Mal. is near the Serpukhovian Stage base, as it is established in the Zabor'e section, appears to be valid.

Finally, I distinguished in the Zabor'e section the new form *Janischewskina adtarusia* Gibshman (msc.) (=*Janischewskina isotovae* Leb. in Mizuno *et al.*, 1997; non *Janischewskina isotovae* Leb. in Grozdilova *et al.*, 1975; Sultanaev *et al.*, 1978; Ueno *et al.*, 1993) in the *Eostaffellina "protvae"* Zone of the Protva Horizon (see Plate III, fig. 20). It is identical to form from the Hina Limestone Group of southwestern Japan that has been pictured by Mizuno and Ueno (1997) in connection with characterization of changes in assemblages of conodonts and foraminifers near the mid-Carboniferous level. In that group, stratigraphic range of *Janischewskina isotovae* Leb. coincides with that of *Cavusgnathus unicornis* (samples H56–H62,5; Tables 1 and 2) and spans the *Gnathodus bilineatus* and a lower half of the *Gnathodus bilineatus–Declinognathodus inaequalis* zones, i.e., it is very close to the species distribution range in the Zabor'e section.

At the same time, holotype of *Janischewskina isotovae* Leb. from the upper Visean Ladeinaya Horizon (exposure Ladeinaya 11, Bed 48, Sultanaev *et al.*, 1978) was described for the first time (Grozdilova *et al.*, 1975) based on a single tangential section. This complicates its correct identification.

CONCLUSION

The studied foraminiferal assemblages from the Zabor'e Quarry are of a high diversity and include, in addition to species crossing the Visean–Serpukhovian boundary, a series of typical Serpukhovian taxa known from the Moscow syneclise, Donetsk basin, Uralian sections, and the upper Chesterian sequence of North America. The identified Serpukhovian forms can be divided in two groups.

The first group includes *Pseudoendothyra globosa* Ros., *Eostaffellina decurta* (Rauser), and *Eostaffellina protvae* (Rauser), all of a narrow stratigraphic and, probably, geographic ranges. These are index species of local foraminiferal zones. Their succession in the Zabor'e section and composition of associated forms show that the early version of foraminiferal zonation in the Serpukhovian Stage (Lipina and Reitlinger, 1970) is quite realistic.

The second group consists of *Eostaffellina paraprotvae* (Rauser), *"Millerella" tortula* Zeller, *Millerella pressa* Thompson, *Janischewskina delicata* Mal., *Janischewskina adtarusia* sp. nov., *Čepekia čepeki* Vasicek et Ruzicka, *Planoendothyra alyutovica* (Reitl.), *"Eostaffella" mirifica* Brazhn., *Loeblichia minima* Brazhn. *Rectoendothyra latiformis*, and Asteroarchaediscidae of the generic lineage *Asteroarchaediscus–Neoarchaediscus–Brenckleina.* These cosmopolitan taxa of a narrow stratigraphic range are widespread in different regions.

The listed taxa are typical of local foraminiferal zones, boundaries between which are defined at the first occurrence levels of *Eostaffellina decurta, Eostaffellina protvae* (the *Eostaffellina* generic lineage) and short-lived species *Pseudoendothyra globosa.* The established zones open a perspective for the interregional correlation of the Serpukhovian Stage based on two generic lineages. The first one represented by *"Millerella" tortula* Zeller *Millerella pressa* Thompson, and short-lived *Brenckleina rugosa* Brazhn. (the *Millerella* lineage) is useful for determination of the Visean–Serpukhovian boundary and for correlation between the studied and coeval deposits of North America. The second *Janischewskina* lineage (*Janischewskina typical* Mikh.–*J. delicata* Mal.–*J. adtarusia* sp. nov.) can be recommended for correlation of the Russian platform sections with those of the southern Urals and Japan.

Thus, foraminiferal species formerly unknown from and identified for the first time in the type area of the Serpukhovian Stage enhance possibilities of interregional correlation of the stage and confirm its global extent.

ACKNOWLEDGMENTS

I am grateful to reviewer A.S. Alekseev for his valuable comments to the manuscript. Foreign colleagues P. L. Brenckle, B. Mamet, J. R. P. Ross, C. A. Ross, L. Hance, M. Viaggi, J. R. Groves, and K. Ueno are greatly appreciated for their advices and interest to my results presented at the poster session of the 14th International Congress on Carboniferous–Permian Stratigraphy and Geology, August 1999, Calgary (Canada). I am also thankful to I.S. Barskov, S.V. Nikolaeva, E.I. Kulagina, and P.B. Kabanov who helped me to select materials for this study. The work was supported by the Russian Foundation for Basic Research, project no. 99-05-65476.

Reviewer A.S. Alekseev

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Stratigraphy and Geological Correlation, Vol. 11, No. 1, 2003, pp. 61–73. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 11, No. 1, 2003, pp. 64–77.
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Correlation of Sections of Lower–Middle Jurassic Deposits in the Greater Caucasus

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Received November 15, 2000

Abstract—Composite stratigraphic sections of Lower–Middle Jurassic (pre-Callovian) deposits, which have been studied in the northern, axial, and southern parts of the Greater Caucasus central segment are described. Stratotypes and fauna of formations and subformations distinguished in these three areas are characterized. The established sedimentation stages characteristic of the whole Jurassic basin of the Greater Caucasus are used to correlate stratigraphic subdivisions of the northern area, where their paleontological content is well studied, with corresponding units of the axial and southern parts, where paleontological data are either insufficient or unknown. As a result, ages of subdivisions in the axial and southern areas are elucidated with a great accuracy at the substage and zonal levels, although previously they have been regarded as units of a wide stratigraphic range, and some researchers even considered them as Paleozoic–Triassic but not Jurassic subdivisions. The relationship between Middle Jurassic (Aalenian–Bathonian) and overlying Callovian–Upper Jurassic deposits has also been elucidated in sections of the northern, axial, and southern area.

Key words: Greater Caucasus, Lower and Middle Jurassic, formations, subformations, correlation, stages, and zones.

INTRODUCTION

The subdivision and dating of Jurassic sequences, highly dislocated, thick, compositionally uniform and poorly characterized by fauna in the axial and southern slope areas of the Greater Caucasus is one of the most complicated problem. Stratigraphic units distinguished here are of debatable ages. Suffice it to say that some of them have been regarded by well-known experts (Afanas'ev, 1956; Afanas'ev *et al.*, 1975; Mel'nikov, 1967; Mel'nikov *et al.*, 1977; Edilashvili *et al.*, 1979) as Paleozoic or Triassic but not the Jurassic formations. Debatable as well is the age relationship between the upper part of sequences and the Upper Jurassic.

The only possible way to solve the problem is to correlate the Lower–Middle Jurassic sequences of axial part and southern zones of the Greater Caucasus with those of northern slope areas, where they are subdivided in detail and well characterized by fauna. Because of deficiency of paleontological remains in the south, the correlation can be based on distinguishing those formations in northern and southern areas, which correspond to particular phases (sedimentation cycles) in geological evolution of both regions and can be subsequently correlated with concurrent cyclothems of the whole Greater Caucasus basin. Works of this kind have been carried out in the western (Panov and Prutskii, 1983) and central Caucasus (Zesashvili, 1970), but the most favorable is a transect across the central Caucasus along the Georgian Military Road. Rengarten (1932) elaborated a scheme of Lower–Middle Jurassic stratigraphic subdivisions for the axial and southern transect segments, which remains valid to date. Ages of formations distinguished by him have not been established however with a proper accuracy and remain debatable so far. A the same time, rock sequences of the northern, axial, and southern areas, which belong to three different structural–facies zones (Panov, 1988, 1995), are closely spaced in this narrowest part of the Greater Caucasus, thus being most suitable for correlation. Exactly here, I analyzed their stratigraphy and correlation based the principle mentioned above. Callovian deposits have been excluded from consideration, since in the Caucasus they are separated from underlying Lower–Middle Jurassic strata by the angular unconformity and represent a single stratigraphic complex with overlying Upper Jurassic deposits that has to be considered separately. In this upper portion, there are intrinsic problems different from those discussed in this work.

All identifications of ammonites mentioned in the text and lacking reference to literary sources have been done by V.P. Kazakova. Descriptions and images of ammonites are available from works by Kazakova (1971, 1984, 1987), Krymgol'ts (1961), Topchishvili (1990), Khimshiashvili (1975).

THE DIGIRIA–OSETIA ZONE

The northernmost zone corresponds to the northern slope of the Liassic–Aalenian basin of the central Caucasus (Fig. 1). Lower to Middle Jurassic deposits over-

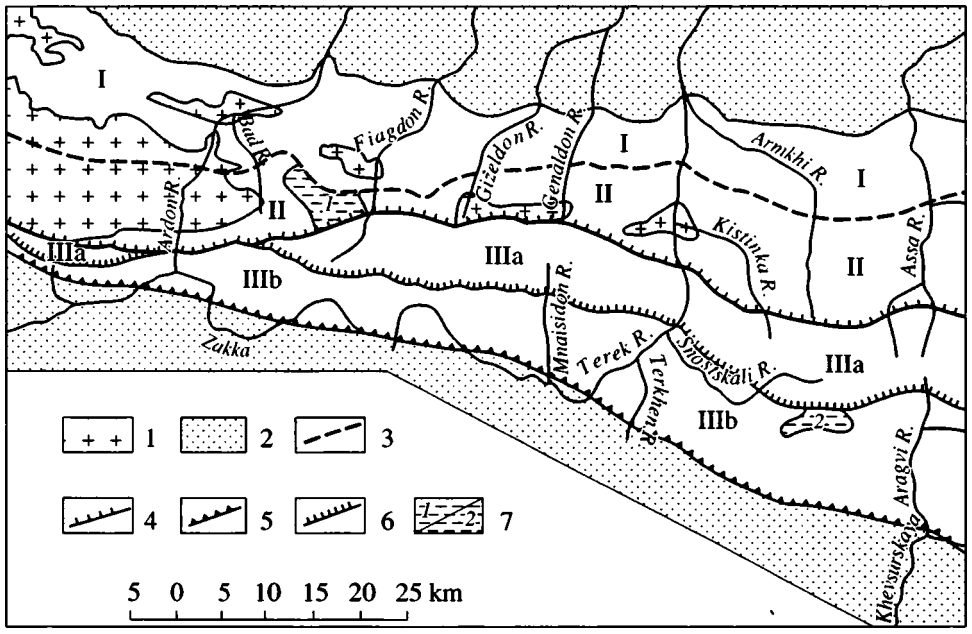


Fig. 1. Scheme of the Early–Middle Jurassic structural–facies zones in the central part of the Greater Caucasus: (1) pre-Jurassic rocks; (2) Callovian, Upper Jurassic, Cretaceous, and Cenozoic rocks; (3) Buron–Lars (Nelkh) fault; (4) Shaukhokh overthrust; (5) Tib overthrust; (6) other thrust faults; (7) Fiagdon (*1*) and Chaukhi (2) nappes; (I) Digoria–Osetia, (II) Pseashkho, and (III) Bzyb–Kazbek zones with exposures (a) of the Tsiklauri, Kazbek, and Gudoshaur formations, and (b) of the Kazbek, Gudoshaurskaya, and Bursachili formations.

lap the Paleozoic basement transgressively, with the angular unconformity. They are composed of shallow-water marine deposits, distinctly stratified and well characterized by fauna (Panov, 1997b). The deposits are subdivided by nondepositional hiatuses into three stratigraphic complexes: the Sinemurian–lower Pliensbachian (the Sadon Formation), the upper Pliensbachian–Toarcian–Aalenian (the Mizur Formation and the Ardon Group), and the Bajocian–Bathonian (the Zgid Formation).

The Sadon Formation (Sinemurian–lower Pliensbachian) is a facies-variable sequence of quartz gritstone and inequigranular sandstone intercalated with coaly, sometimes graphitic mudstone and siltstone, andesitic and dacitic lavas, their tuffs, tuffites, and other volcanogenic-sedimentary deposits. All these rocks occur as alternating lenticular strata, with volcanites dominating in northern sections and gradually giving way in the southward direction to terrigenous rocks, which make up 90–95% of the formation strata near the southern boundary of the zone. The Sadon Formation is replaced by the Kistinka Formation in a more southerly Pseashkho zone (see below). The thickness of the Sadon Formation is up 700–800 m at the southern boundary of the zone and sharply reduces northwards, where the formation is completely pinching out in northern exposures along the Urukh River. The Early Jurassic age of the Sadon Formation is established based on plant fossils (determinations by A.N. Krishtofovich), and its correspondence to the early–middle Lias is based on identifications of pelecypods and brachiopods (V.F. Pchelintsev and G.T. Pchelintseva, A.S. Moiseev). According to the conformable boundary with upper Pliensbachian deposits near southern limits of the zone and to correlation with formations of analogous lithology and stratigraphic position in southern areas (Panov and Gushchin, 1987), the Sadon Formation is referred to the Sinemurian–lower Pliensbachian.

The Mizur Formation (upper Pliensbachian) rests with erosional hiatus on the Sadon Formation and, if the latter is missing, transgressively overlaps the Paleozoic basement. In its typical sections, the Mizur Formation is divided in three subformations. The lower subformation (0–75 m) is composed of sandstone and siltstone enclosing fine pebble and gritstone interlayers with plant fossils and with shelly limestone lenses in places. In the middle subformation (40–115 m), siltstone and mudstone with siderite concretions are intercalated with limestone interlayers bearing *Amaltheus* spp. The thickest upper subformation (115–250 m) is composed of fine-grained platy sandstone with *Amaltheus* spp. (Fig. 2, Section 1). In the southward direction, sandy material is replaced by clayey–silty material, and the formation looses its three-unit structure to become a single sequence of sandy–clayey siltstone with interlayers of fine-grained sandstone. Near the southern boundary of the zone, where it conformably rests on the Sadon Formation, the unit is 600 m thick. South of the Buron–Lars fault, the Sadon Formation is replaced by the lower part of the Tsiklauri Formation, an inherent

unit of the Pseashkho zone (see below). The direct transition from the Sadon to Kistinka Formation and from the Mizur Formation to the lower part of the Tsiklauri Formation has been described in the Baron–Lars fault zone at the Ardon River right bank (Avdonin, 1963).

The Ardon Group is represented by a uniform complex of alternating clayey–silty and sandy–silty–clayey deposits. It conformably rests on the Mizur Formation. The group is subdivided into four regressive sedimentary rhythms. The first (Galiat and Ksurt formations), second (Dzhimara Formation), and third (Salgi and Koirakh formations) rhythms are made up of clayey–silty sequences, gradually giving way in the upward direction to sandy–silty deposits. The forth rhythm is incomplete, composed of clayey–silty beds of the Fortaukh Formation. Sandy–silty laminae appearing in upper parts of the rhythms are related in origin to progradation of distal parts of the Avar submarine fan located easterly, in the Agvali–Khiv zone, into the territory of the Digoria–Osetia zone (Panov, 1997b).

The Galiat Formation (the lowermost lower Toarcian) is composed of black mudstone with siderite concretions and with rare siltstone and sandstone interlayers bearing *Dactylioceras tenuicostatum* J. et B., *D. semicelatum* Simps., *Hidaites* cf. *serpentinum* Rein., and *Harpoceratoides* cf. *alternatus* Simps., which are characteristic of the *D. tenuicostatum* Zone and of the lower *H. falcifer* Zone. The thickness range of the Galiat Formation is 120–480 m.

The Ksurt Formation (upper lower Toarcian) is largely composed of clayey siltstone with frequent interlayers and beds of fine-grained sandstone and with numerous members of alternating sandstone and siltstone. The rocks yield *Harpoceras* cf. *falcifer* Sow., *Hildaites* aff. *levisoni* Simps., *Hildoceras* cf. *bifrons* Brug., and *Dactylioceras commune* Sow., characteristic of the upper part of the *H. falcifer* Zone coupled with the *H. bifrons* Zone. The unit thickness increases from 300 m in the northwest to 750 m in the southeast.

The Dzhimara Formation (upper Toarcian), represented by a sequence of mudstone and banded sandy–clayey siltstone, is crowned by the member (100–170 m thick) of alternating fine-grained sandstone, siltstone, and mudstone beds. The total thickness of the formation increases from 160 m in the west to 650 m in the east. Ammonites Phymatoceratinae, *Grammoceras* ex gr. *thouarsense* Orb., *Pseudogrammoceras fallaciosum* Bayle, and *Dumortieria* cf. *gundershofensis* Buckm., which have been found in the rocks, imply that all three upper Toarcian zones are distinguishable in the formation.

The Salgi Formation (the uppermost interval of the upper Toarcian–lower Aalenian *D. levesquei* Zone) is composed of black mudstone and thin-banded clayey siltstone. The unit is 320–1000 m thick. Late Toarcian *Dumortieria* sp., *Pleydellia* sp. still occur in its base, and the main part of the formation encloses early Aalenian *Leioceras opalinum* Rein.

The Koirakh Formation (terminal lower–initial upper Aalenian) includes members of black mudstone with siltstone interlayers and members of frequently alternating fine-grained sandstone, siltstone, and mudstone. The lower part of the unit yields *Leioceras* ex gr. *comptum* Rein., and *Tmetoceras* cf. *scissum* Ben., while the upper one bears *Staufenia* spp. and *Ludwigia murchisonae* Sow. According to its paleontological content, the formation corresponds to the upper part of the *L. opalinum* Zone coupled with the lower part of the *L. murchisonae* Zone. The formation is 200 m thick in the west to 1100 m thick in the east.

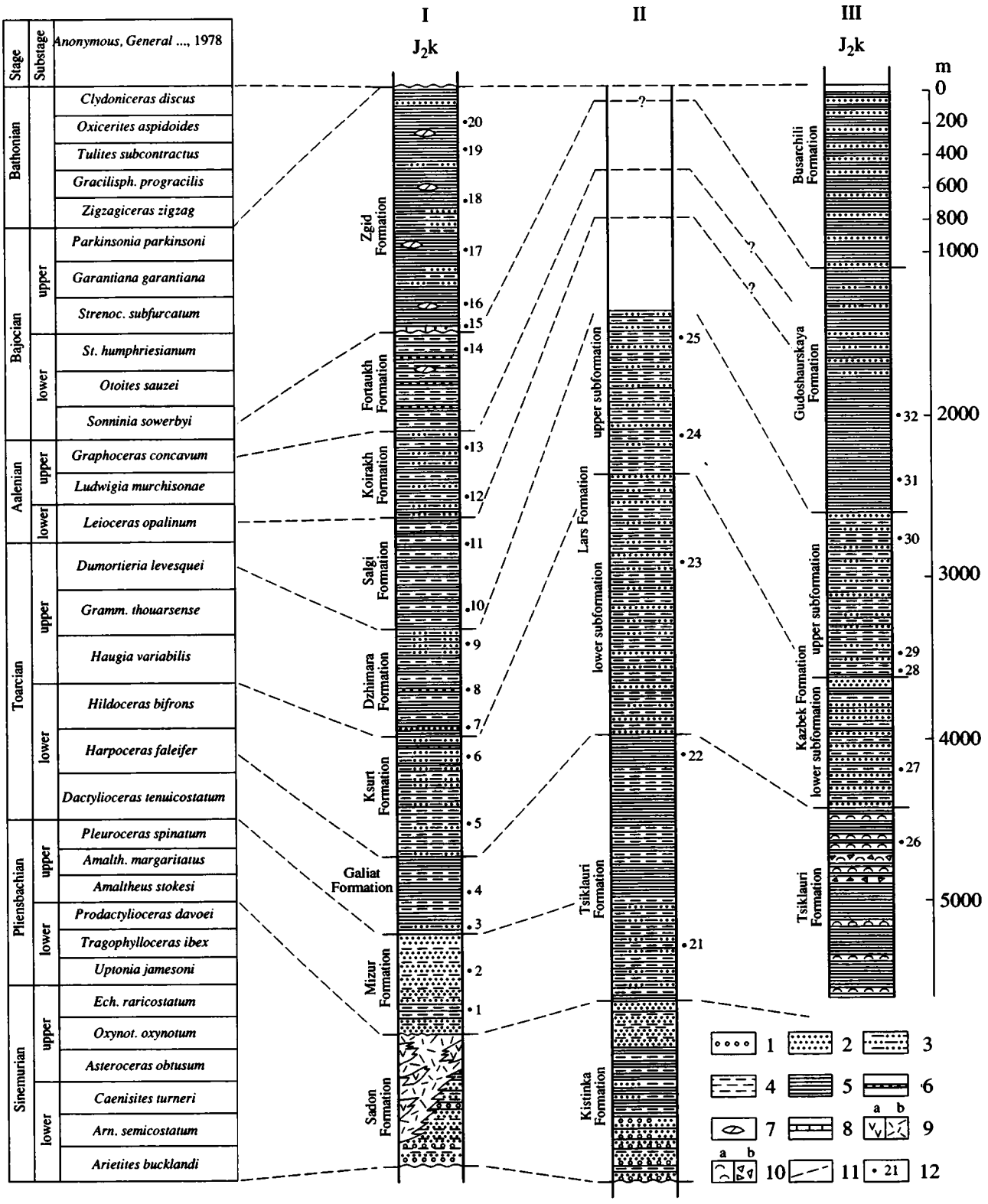
The Fortaukh Formation (upper Aalenian–lowermost Bajocian) is represented by a uniform sequence of mudstone and clayey siltstone, which are 250–750 m thick in total. *Staufenia* spp., *Graphoceras concavum* Sow., and *Tatrophyloceras subglobosum* Besn. occurring in the unit enable recognition of the *L. murchisonae* and *Gr. concavum* zones, and of lower beds of the *S. sowerbyi* Zone.

The Zgid Formation (Bajocian) with the basal horizon of bioclastic limestone or massive sandstone with pebbles transgressively overlaps different beds of the Fortaukh Formation. This is a uniform sequence of mudstone and clayey siltstone, which enclose lenticular interbeds of marly siltstone with large bun-shaped siderite nodules. In places, rather thick sandstone members occur at several levels. In particular localities, the Zgid Formation is of a highly variable thickness (from 1500 m to 0 m) owing to development of the pre-Callovian erosional hiatus. Found ammonites *Witchellia sayni* Haug., *Sonninia* sp., *Stephanoceras humphriesianum* Sow., *Partschiceras abichi* Uhl., *Leptosphinctes* spp., *Garantiana* ex gr. *alticosta* Wetz., *Parkinsonia rarecostata* Buckm., and *P. parkinsoni* Sow. (Beznosov, 1967) suggest that all the lower and upper Toarcian zones are distinguishable in the formation. Sequences of the Digoria–Osetia zone are lacking the Bajocian deposits.

The lower Callovian transgressive deposits with fauna overlie the Zgid Formation with a hiatus and, in places, with the angular unconformity (up to 80°). Available opinions (Beznosov, 1967; *Yura Kavkaza*, 1992) that the Targim Formation at the base of these deposits is of the middle–early Bathonian age are misleading, since this unit bears ammonites of the early Callovian age (*Ob'yasnitel'naya zapiska...*, 1973) but none of the Bathonian forms has ever been found.

THE PSEASHKHO ZONE

This zone corresponding to the modern axial structure of the Greater Caucasus was distinguished for the first time in the northwestern region (Panov and Prutskii, 1983). In the Terek River basin, it lies between the Buron–Lars fault in the north and the Shaukhokh fault in the south (Fig. 1). Rengarten (1932) who was the first to establish the last fault at the right bank of


Stage
Substage
Anonymous, General ..., 1978
Bathonian
Clydoniceras discus
Oxicerites aspidoides
Tulites subcontractus
Gracilisph. progracilis
Zigzagiceras zigzag
Bajocian
upper
Parkinsonia parkinsoni
Garantiana garantiana
Strenoc. subfurcatum
lower
St. humphriesianum
Otoites sauzei
Sonninia sowerbyi
Aalenian
upper
Graphoceras concavum
Ludwigia murchisonae
lower
Leioceras opalinum
Toarcian
upper
Dumortieria levesquei
Gramm. thouarsense
Haugia variabilis
lower
Hildoceras bifrons
Harpoceras faleifer
Dactylioceras tenuicostatum
Pliensbachian
upper
Pleuroceras spinatum
Amalth. margaritatus
Amaltheus stokesi
lower
Prodactylioceras davoei
Tragophylloceras ibex
Uptonia jamesoni
Sinemurian
upper
Ech. raricostatum
Oxynot. oxynotum
Asteroceras obtusum
lower
Caenisites turneri
Arn. semicostatum
Arietites bucklandi
I
J2k
II
III
J2k
m
0
200
400
600
800
1000
2000
3000
4000
5000
Zgid Formation
Fortaukh Formation
Koirakh Formation
Salgi Formation
Dzhimara Formation
Ksurt Formation
Galiat Formation
Mizur Formation
Sadon Formation
Busarchii Formation
Gudoshaurskaya Formation
upper subformation
Lars Formation
lower subformation
Kazbek Formation
upper subformation
lower subformation
Tsiklauri Formation
Tsiklauri Formation
Kistinka Formation
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12

Terek River showed it in the geological map and profile under the name the Kazbek overthrust bounding on the south the "Dar'yal subzone" thrust over the southerly "Kazbek–Gudoshaurskaya zone." Rengarten reasoned that the "tuff, porphyry, and schists of the Tsiklauri Formation" are thrust over the Kazbek Formation. According to my observations, deposits of the Tsiklauri Formation of the northern type are thrust, together with the overlying Lars Formation (see below), over the southern-type rocks of the Tsiklauri Formation. After geological prospecting (Enna, 1991), the overthrust was traced further westward up to Fiagdon River, now under the name Shaukhokh fault, as the boundary between the Pseashkho (Adaikhokh–Dar'yal after Enna) and Bzyb–Kazbek structural–facies zones (Enna attributed the latter to central and southern slope zones). The zones greatly differ in tectonic deformations and structure of Lower Jurassic sequences.

Lower Jurassic deposits of the Pseashkho zone overlie transgressively (but usually without the angular unconformity) different horizons of Upper Carboniferous and Permian deposits. Characteristic of them are the large thickness of continuous section and postdiagenetic alterations of rocks at the proto- and meso-metamorphic stages (Chichua *et al.*, 1985), which led some researchers to conclude that the rocks are of Paleozoic–Triassic age. These intensely dislocated rocks are intruded by numerous transverse diabase dikes. The section of Lower Jurassic deposits is subdivided (Panov, 1995; *Yura Kavkaza*, 1992) into the conformable Kistinka, Tsiklauri, and Lars formations (Fig. 2, Section II). As is mentioned above, two lower formations have been referred by some researchers to the Paleozoic–Triassic.

The Kistinka Formation (Sinemurian–lower Pliensbachian) distinguished by Rengarten (1932) is basal one on the Jurassic sequence. It is composed of facies-variable members and beds of quartzitic sandstone and gritstone alternating with members of clayey–silty, often coaly shale with plant fossils. Rengarten did not suggested the formation type section but mentioned that its sediments are widespread in the Kistinka River valley. The Skaron section at the right bank of Midagrabindon River (name of Gizel'don River in the upper reaches), where rocks of the Kistinka Formation perfectly exposed and bear fauna, can be regarded as the lectostratotype. In this area, the karstified surface of Permian limestones is overlain, with traces of erosional hiatus, by the following rock units:

(1) Alternation of lenticular beds of inequigranular quartzitic sandstone and cherty quartz gritstone; unit is 5–6 m thick.

(2) Black coaly clayey–silty shale with lenticular interlayers of inequigranular quartzitic sandstone and gritstone; abundant plant remains and pelecypods *Chlamys* ex gr. *textoria* Schl., *Meleagrinella* aff. *neglecta* Moberg, and *Oxytoma* ex gr. *inaequivalvis* Sow. are characteristic of the unit that is 25–30 m thick.

(3) Massive inequigranular quartzitic sandstone with gritstone interlayers; unit is 35–40 m thick.

(4) Alternating beds (2–20 m) of black clayey–silty shale and gray fine-grained quartzitic sandstone: rocks bear pelecypods *Aequipecten* cf. *priscus* Schl., *Avicula yaltensis* Moiss., and *Cardinia* aff. *copides* Ryckh.; unit is 150–165 m thick.

(5) Members of massive coarse-grained quartzitic sandstone with gritstone interlayers alternating with subordinate members of coaly silty shale also with gritstone lenses; unit is 150 m thick.

(6) Regularly alternating members of fine-grained quartzitic sandstone and black silty shale with pelecypods *Parallelodon similis* Terq. and *Lima* cf. *exaltata* Terq; unit is 225–250 m thick.

(7) Black clayey and banded silty shales with beds and numerous members (up to 5–10 m) of fine-grained quartzitic sandstone, which are concentrated near the base; unit is 220–225 m thick.

(8) Alternating beds of massive fine-grained quartzitic sandstone and banded silty shale; members of mas-

Fig. 2. Correlated sections of Lower–Middle Jurassic deposits of the central Caucasus: (1) conglomerate and gritstone; (2) sandstone; (3) sandy–silty–clayey alternation; (4) siltstone; (5) mudstone; (6) concretion conglomerate; (7) lenticular beds and concretions of marly siltstone; (8) limestone; (9) andesite–dacitic lavas (a) and pyroclastic rocks (b); (10) basalt lavas (a) and pyroclastic rocks (b); (11) boundaries between formations and subformations; (12) localities of found ammonites. (I–III) Stratigraphic sections of (I) Digoria–Osetia, (II) Pseashkho, and (III) Bzyb–Kazbek zones. Occurrence levels of ammonites (1) *Amaltheus margaritatus* Montf., *Am. (Pseudoamaltheus) laevis* Qu.; (2) *Amaltheus margaritatus* Montf., *Am. depressus* Simps., *Am.* cf. *stokesi* Sow., *Am. (Pseudoamaltheus) laevis* Qu.; (3) *Hildaites* cf. *serpentinum* Rein,; (4) *Dactylioceras semicelatum* Simps., *D.* cf. *tenuicostatum* J. et B.; (5) *Harpoceras* cf. *falcifer* Sow., *Hildaites* aff. *levisoni* Simps.; (6) *Dactylioceras commune* Sow., *Hildoceras* cf. *bifrons* Brug.; (7) family Phymatoceratinae; (8) *Grammoceras* ex gr. *thouarsense* Orb., *Pseudogrammoceras fallaciosum* Bayle; (9) *Dumortieria* cf. *gundershofensis* Buckm.; (10) *Pleydellia mactra* Dum., *Dumortieria* sp.; (11) *Leioceras opalinum* Rein., *L.* cf. *costosum* Qu., *L.* cf. *gotzendorfensis* Dorn.; (12) *Leioceras* ex gr. *comptum* Rein., *Tmetoceras* cf. *scissum* Ben.; (13) *Staufenia (Costileioceras) sinon* Bayle, *St. (C.) opalinoides* Mayer, *St. (C.)* cf. *sehndensis* Hoffm., *Ludwigia murchisonae* Sow.; (14) *Graphoceras* cf. *concavum* Sow., *Tatrophylloceras subglobossum* Besn.; (15) *Witchellia sauni* Haug; (16) *Sonninia* sp., *Emileia* sp.; (17) *Stephanoceras humphriesianum* Sow.; (18) *Partschiceras abichi* Uhl., *Leptosphinctes* spp.; (19) *Garantiana* ex gr. *alticosta* Wetz.; (20) *Parkinsonia rarecostata* Buckm., *P. subarietis* Wetz., *P. parkinsoni* Sow.; (21) *Amaltheus* cf. *laevigatus* How.; (22) *Harpoceratoides* sp.; (23) *Polyplectus* sp.; (24) *Haugia variabilis* Orb.; (25) *Dumortieria* cf. *munieri* Haug.; (26) *Amaltheus subnodosus* J. et B., *Am. striatus* How. (in Kakhetia, along Stori River); (27) *Dactylioceras* sp., *Nautilus austracoides* J. et B.; (28) *Haugia* sp., *Denckmannia* sp.; (29) *Grammoceras thouarsense* Orb.; (30) *Dumortieria* cf. *bleicheri* Ben., *Pleydellia* cf. *subcompta* Br.; (31) *Pleydellia mactra* Dum.; (32) *Leioceras opalinum* Rein.

sive sandstone up to 20 m thick and rare lenticular gritstone interlayers are also encountered. The unit is 215–225 m thick.

At the top, the Kistinka Formation is overlain by shales of the Tsiklauri Formation. In the stratotype and other complete sections, the formation is of a three-unit structure. Its lower part (400 m thick along Midagrabindon River, beds 1–5, more than 250 m thick along Kistinka River) reveals prevalence of coarse-grained sandstone and gritstone; members of coaly silty shale, often rich in scattered quartz gravel ("plum-cake rocks") are also typical of this interval. The middle part (420 m) is predominantly composed of silty rocks, although in the Midagrabindon River section (beds 6–7) it also includes members of quartzitic sandstone fine-grained in this case. The upper part (225 and 330 m correspondingly) is mainly composed of members of fine-grained quartzitic sandstone and subordinate members of silty shale. Quarts grains represent 95% of clastic material in sandstone and gritstone of the lower formation part, whereas sandstone of the upper part consists of quartz (about 40%) and plagioclase (15%) grains mixed with rock fragments. In some places (exposures along the Midagrabindon, Genaldon, Terek rivers), lenses of acid volcanogenic rocks (albitophyre lavas and tuffs) are present among sedimentary strata. The total thickness of the Kistinka Formation is about 1000–1100 m. The abundant quartz and kaolinite present in the rocks suggest that deposits of the Kistinka Formation originated by washout and redeposition of the pre-Jurassic weathering crust (Ivanov, 1963).

Apart from the stratotype area, normal stratigraphic contacts between the Kistinka Formation and underlying rock complexes have been observed in head areas of the Bad, Tsariitdon (right tributary of the Fiagdon River), and Genaldon rivers, where it transgressively overlies either the various Permian limestone strata or the Upper Carboniferous quartzite and siltstone beds. According to my observations along the Terek River, stratigraphic contacts are demolished here, and rocks of the Kistinka Formation are in tectonic contacts with Paleozoic granitoids of the Dar'yal and Gvileti massifs and with fragments of Upper Carboniferous deposits enclosing the latter. According to geological survey data (Kipiani, 1984), the Kistinka Formation transgressively overlies the denuded surface of granitoids even here, and its gritstone layers include rare pebbles of plagiogranite similar to that from the Dar'yal massif (Kipiani, 1984; Chikhradze, 1975). The Kistinka Formation is conformably overlain everywhere by clayey–silty rocks of the upper Pliensbachian Tsiklauri Formation.

These data alone indicate that the Kistinka Formation does not belong to the Paleozoic–Triassic complex and represents a basal subdivision of the Jurassic sequence. Preliminary data on Late Permian brachiopods (the Mt. Gurgala, the Dar'yal massif arch) and tabulates (upper reaches of Bad River) found in the lower part of the Kistinka Formation (Mel'nikov *et al.*, 1977) have not been supported in subsequent studies. Chikhradze (1975) suggested that these fossils were collected most likely from Permian limestone and marl beds underlying the Kistinka Formation. Information about the Early Triassic ammonite *Arctoceras* sp. found at the right bank of Kistinka River and Triassic crinoids from the right bank of Midagrabindon River (Mel'nikov *et al.*, 1977) is doubtful as well, because crinoids were identified at the family level only (!), and unidentifiable ammonite just resembles *Arctoceras* (?) sp. (*Anonymous, Annual...*, 1960).

In the composition, structure, and stratigraphic position, the Kistinka Formation is similar to the Sadon Formation of the Digoria–Osetia zone described above. The facies transition from one formation to another is observable in the Bad and Tsariitdon river valleys. The age similarity between formations is evident from paleontological data as well. Pelecypods found in the Kistinka Formation section along Midagrabindon River (see above) are characteristic (according to Pchelintseva and Sibiryakova who identified them) of the lower and middle Lias. According to Topchishvili who identified the pelecypod form *Lima* cf. *hausmanni* Dum. in gritstone sample from the lower part of the Kistinka Formation at the periphery of the Dar'yal massif, it is also typical of the lower and middle Lias. The Kistinka Formation conformably overlain by Upper Pliensbachian deposits with fauna is correlative with the Svanetia and Stori formations of the southern areas because of lithologic similarity and stratigraphic position (Panov and Gushchin, 1987). The Sinemurian–early Pliensbachian age has been established for the last two formations based on paleontological data, and the Kistinka Formation can be attributed to the same stratigraphic interval.

The Tsiklauri Formation (upper Pliensbachian–lowermost lower Toarcian) conformably, though with a sharp lithological contact, rests on the Kistinka Formation. Rengarten (1932) distinguished under the same name a sequence, "... in which dark shale dominates over more rare and thin interlayers of sandstone and quartzite." He also conceived that characteristic of the formation are "lava sheets of diabase and porphyrite." My Georgian colleague (Chikhradze, 1979) and me, as well as the results of geological mapping (Pis'menskii and Snezhko, 1991), showed that magmatic bodies of the Pseashkho zone represent dikes intruding the folded structure, and hence they should be excluded from stratigraphic characterization of the formation. This peculiarity discriminates the Tsiklauri Formation of that northern zone from the synonymous formation of the southern Bzyb–Kazbek zone (see below), where synchronous volcanic complexes do occur.

The stratotype of the Tsiklauri Formation has not been suggested by Rengarten as well. The section at the slope of Mt. Molchekort, the right bank of Kistinka River, can be considered as its lectostratotype. Sand-

stones of the Kistinka Formation are conformably, though with a rather sharp lithological boundary, overlain here by the following units:

(1) Massive banded silty and sandy–silty shales with lenticular interlayers of fine-grained sandstone; unit is 50 m thick.

(2) Thin-banded clayey–silty shale with interbeds of massive silty shale; unit is 400 m thick.

(3) Massive banded silty shale with subordinate interbeds of dark silty and sandy–silty shales; unit is 200 m thick.

(4) Black "silky" shale with interlayers of thin-banded silty shale; unit is 300 m thick.

(5) Thin-banded silty shale with subordinate interlayers of coarse-banded silty shale; unit is 700 m thick.

These units are conformably overlain by sandy–clayey to silty sediments of the Lars Formation.

In this section, like in some others, the Tsiklauri Formation is subdivided in two parts interconnected via gradual transition. The lower part (about 650 m) is mainly composed of massive banded–interlensing silty and sandy–silty shales, and the upper one (1000 m) is represented by black clayey and thin-banded silty shales.

In the Terek River valley north of the Dar'yal massif, the Tsiklauri Formation is fragmentary exposed in faulted blocks, and its apparent thickness is here only 650 m, though the lower (siltstone with sandstone interbeds) and upper (mainly clayey) parts of the formation can be easily discriminated even in this area. This fact gave grounds to Mel'nikov (1967) to distinguish, in the Terek–Assa interfluve, the lower siltstone part as the independent Belaya Reka Formation that reduced the scope of the Tsiklauri Formation to its upper clayey–silty part. Further westward, in the Terek–Ardon interfluve, the Tsiklauri Formation is known in structure of several tectonic sheets separated by faults. In places, where stratigraphic contacts are preserved, on can see conformable, though sharp boundaries of this subdivision with the underlying Kistinka and overlying Lars formations. In the mentioned sheets, the Tsiklauri Formation as a whole represents a uniform sequence of black shales or thin-banded silty shales intercalated with several thin (up to 1–3 m thick) horizons of schistose volcanic rocks (?) and basic tuffs. Judging from available data, the total thickness of the formation is about 1500 m.

The Tsiklauri Formation age is established at present rather precisely. Its stratigraphic position denies the unit affiliation with the Paleozoic–Triassic, as it was formerly suggested (Afanas'ev, 1956; Afanas'ev *et al.*, 1975; Edilashvili *et al.*, 1979). The lower, mainly siltstone part of the formation is similar in composition, structure, and stratigraphic position to the Mizur Formation of the southern Digoria–Osetia zone, and represents the facies analogue of the latter in the Bad River basin. Hence, both correlative subdivisions span the upper Pliensbachian interval (Fig. 2). This accords well with known paleontological data: *Amaltheus* cf. *laevigatus* How. was found in the lower part of the Tsiklauri Formation at the Terek River bank along with *Fuciniceras* cf. *boscense* Reyn. that is also known from the Olgit gorge at the left bank of Armkha River (Zesashvili, 1979). In addition, *Amaltheus* sp. and *Calliphylloceras* cf. *bicicolae* Men. were found at the Assa River bank near the Nelkh River mouth, its right tributary (Topchishvili, 1988). The upper, mainly clayey part of the Tsiklauri Formation can be reliably correlated with the Galiat Formation of the Digoria–Osetia zone, which is similar in the composition and stratigraphic position. The latter spans the basal lower Toarcian interval (from *D. tenuicostatum* Zone to lower beds of *H. falcifer* Zone). The additional finding of *Harpoceratoides* sp. is known from stratigraphic level above that of late Pliensbachian ammonites at the Assa River bank.

The Lars Formation (uppermost lower Toarcian–upper Toarcian) was distinguished for the first time by Robinson (1948; unpublished) as a sandy–silty–clayey sequence conformably overlying clay rocks of the Tsiklauri Formation. Rengarten described these deposits along with rocks of the Ardon Group of the Digoria–Osetia zone under the name Dzherakh Formation. In the southern part of the region, especially in the Kistinka River basin where rocks of the Lars Formation are crossed by diabase dikes, they were included into the scope of the Tsiklauri Formation.

The Lars Formation is a stratified sequence of prevailing banded sandy–silty and clayey–silty sediments, which are intercalated with numerous sandstone interlayers and beds. It conformably rests on rocks of the Tsiklauri Formation, and its distribution area is bounded by the Buron–Lars fault (the Nelkh fault east of Terek River) on the north. Robinson distinguished the Lars Formation in the Terek River valley between the villages of Verkhnii Lars and Nizhnii Lars, but different parts of the formation are exposed here in a series of tectonic slices inappropriate for the stratotype definition. The section along Khankol River (right tributary of the Armkha River), where deposits of the Lars Formation have a monoclinal structure and bear fauna, can be suggested as their lectostratotype. Rocks of the Tsiklauri Formation are overlain here by the following units:

(1) Coarse-banded sandy–clayey to silty shale with irregular interlayers and separate beds (up to 10–15 cm thick) of fine-grained sandstone. The thickness is 1050 m.

(2) Thin-banded silty shale intercalated with thin siltstone interlayers and rare beds of fine-grained sandstone bearing *Polyplectus* sp. (identified by E.P. Sergeeva); unit is 345 m thick.

(3) Coarsely banded sandy to silty shale; unit is 100 m thick.

(4) Thin-banded silty shale with very rare individual sandstone beds; unit is 720 m thick.

(5) Coarsely banded sandy to silty shale; unit is 170 m thick.

(6) Thin-banded silty shale with *Dumortieria* cf. *munieri* Haug.; unit is 325 m thick.

In this section, like all over the Terek–Armkha–Assa interfluve, the Lars Formation is subdivided in two subformations. The *lower Lars Subformation* (upper lower Toarcian) consists of irregularly alternating coarse-banded members of light-colored sandy and dark clayey siltstone beds (few cm thick), which are intercalated with sandstone interlayers and beds. Some members of thin-banded clayey siltstone include packets (5–15 cm thick), which exhibit frequent flyschoid interbedding of fine-grained sandstone and clayey siltstone. The lower subformation is 1500–1700 m thick. The *upper Lars Subformation* (upper Toarcian) differs from the lower one in the prevalence of fine-banded clayey–silty rocks, although coarse-banded rocks and individual sandstone interlayers are also available there. The apparent thickness of the upper subformation is 850–1200 m. Beside the described section, ammonite *Haugia variabilis* Orb. was found at the left bank of the Terek River.

Further westward, in the Ardon–Terek interfluve, the Lars Formation can easily be also subdivided in two subformations despite the tectonic deformation of exposures resembling those of the Tsiklauri Formation. The lower Lars Subformation is composed there of alternating members (25–70 m) of fine-grained sandstone and sandy siltstone (5–15 m). Its total thickness estimated in different tectonic slices is 350–370 m. The upper Lars Subformation is divisible here in two parts. In the lower one, it represents a uniform rock mass of thin-banded clayey siltstone (550 m), which grade upward into a layered mass of coarse-banded sandy and sandy–clayey siltstone with numerous sandstone interlayers and few limestone beds (775–865 m).

By the structure and stratigraphic position, the lower Lars Subformation can be reliably correlated with the Ksurt Formation of the Digoria–Osetia zone (Fig. 2) and referred to the upper part of the *H. falcifer* Zone coupled with the *H. bifrons* Zone of the lower Toarcian that is indicated by finds of *Polyplectus* sp. The upper Lars Subformation is well correlative, according to its stratigraphic position and lithology (especially in the Ardon–Terek interfluve), with the Dzhimara Formation. It corresponds to the upper Toarcian in its full scope that is evident from found ammonites characteristic of the *H. variabilis* and *D. levesquei* zones.

In the Pseashkho zone of the central Caucasus, both subformations crown the section of Jurassic deposits. All the deposits described above are intensely deformed and penetrated by diabase dikes of the Kazbek complex (Pis'menskii and Snezhko, 1991), which are of the Bajocian age in opinion of many researchers. In the eastern Caucasus, continuation of the Pseashkho zone is represented by the Metlyuta–Akhtychai zone (Panov, 1988), where the Jurassic sequence includes in addition the Aalenian deposits, also deformed and crossed by diabase dikes. At the right bank of Samura River, Aalenian strata are unconformably overlain by the Upper Jurassic beds of the Shakhdag area. The absence of Bajocian and Bathonian deposits in the Pseashkho zone seems to be an inherent feature. Beginning since the initial Bajocian, there was under development the central Caucasian system of uplifts composed of Lower Jurassic–Aalenian deformed deposits intruded by diabase dikes.

THE BZYB–KAZBEK ZONE

The zone extends throughout the southern slope of the Greater Caucasus (Panov, 1988). In its central part, the zone is located between the Shaukhokh fault in the north and the Tib fault (Vardanyants, 1935) in the south (Fig. 1). Along the plane of the Tib thrust fault, the Lower–Middle Jurassic deposits of the Bzyb–Kazbek zone override the Upper Jurassic–Cretaceous flysch of the Chiauri synclinorium. The Bzyb–Kazbek zone corresponds to the axial trough of the Greater Caucasus basin. The continuous sequence of Lower–Middle Jurassic deposits of this zone is composed predominantly of deep-sea clay and flyschoid sediments nearly barren of fossils and hosting intercalations of volcanogenic basic rocks. The deposits are intensely deformed and transformed at the of proto- and meso-metamorphic stages (Chichua *et al.*, 1985).

In the central Caucasus, Rengarten (1932) subdivided Lower–Middle Jurassic deposits of the Bzyb–Kazbek zone, where their base is not exposed, into the Tsiklauri, Kazbek, Gudoshaur, and Busarchili formations. In his opinion, contacts between these formations are tectonic. Lower–Middle Jurassic deposits do form there an imbricated nappe system of the accretionary type. This is a complex of tectonic slices with overthrust planes dipping northwards. Age of deposits constituting the system gradually becomes younger from north to south, and in each sheet sedimentary beds dipping northward occur in a normal stratigraphic succession. In places, particular tectonic slices are composed of two successive formations displaying normal stratigraphic relationships, and it is possible to argue that all the formations distinguished by Rengarten form a single continuous section (Panov, 1995).

The Tsiklauri Formation (upper Pliensbachian–lowermost Toarcian) is basal one in the exposed sequence (Fig. 2, Section III). In contrast to its counterpart in the Pseashkho zone, the subdivision is composed exclusively of clay shales with numerous lava flows of pillow and massive amygdaloidal basalts, lava breccias, spilite, vitrophyre, variolite, and conformable diabase sills representing subvolcanic bodies (Beridze, 1981; Pis'menskii and Snezhko, 1991; Chikhradze, 1979; and my data). Taking into account these distinctions, Enna (1991) subdivided the deposits in question into the Dzamarash and Bugul'ta formations that is hardly correct because exactly here, in the vicinity of the village

of Tsiklauri, Rengarten distinguished the type section of the Tsiklauri Formation. I regard this subdivision as the Tsiklauri Formation of the southern type.

The content of volcanogenic rocks in the Tsiklauri Formation varies in both the stratigraphic and lateral directions. These rocks are most widespread in the Ardon–Terek interfluve (the Fiagdon nappe included) and in the Terek River valley, where lava flows and diabase sills are up to 30–60 m thick in the upper part of the formation and constitute 60–80% of the section. For instance, 24 lava flows and 14 diabase sills, the total thickness of which is 299 m, have been counted in the Terek River valley (Beridze, 1981). The total thickness of sediments hosting volcanic rocks may be up to 450–500 m. Volcanogenic constituents of the Tsiklauri Formation are similarly widespread in the Chaukhi nappe (Rengarten, 1932; Sukhanov and Tsvetkov, 1975). A remarkable exposure of magmatic rocks of the formation, the last one on the east, is situated near the Assa River mouths, in the Chingis-klde Range, where I described a member of basalt flows and diabase sills with minor beds of shale, the total thickness of which is 150 m.

In the southernmost exposures, thick lava flows and diabase bodies pinch out and only individual members (up to 10–40 m) or beds (several meters thick) of foliated tuff and more rare pillow-lava flows are present in the Tsiklauri Formation. This situation decreases distinction between deposits of the Tsiklauri Formation of the southern and northern (for instance, in the Ardon–Terek interfluve) zones and confirms their concurrent age.

The total thickness of the Tsiklauri Formation of the southern type exceeds 1100 m. It is believed that its distribution area is bounded by the Adaikom–Kazbek overthrust on the south (Vardanyants, 1935; Rengarten, 1932), though in some places, where the Tsiklauri Formation is exposed south of it (the Fiagdon River heads, the right-bank area of Snostskali River, the Khevsurskaya Aragvi River heads), the Tsiklauri Formation is conformably overlain by the Kazbek Formation. I reckon the Tsiklauri Formation of the southern zone as concurrent in age with synonymous subdivision of the northern zone, the age of which is precisely established. This is confirmed to some extent by findings of upper Pliensbachian ammonites *Amaltheus subnodosus* Joung et Bird and *Am. striatus* How. (identified by N.G. Khimshiashvili) in shales with horizons of spilitic pillow lava and tuffs (the Tsiklauri Formation of the southern type) in the eastern extension of the Bzyb–Kazbek zone, in the Stori River locality of Kakhetia (Pruidze, 1979).

The Kazbek Formation (terminal lower–upper Toarcian) conformably overlying the Tsiklauri Formation is also conformably overlain by the Gudoshaur Formation. Rengarten (1932) distinguished under this name "the formation of banded shale with frequent interlayers of banded quartzite", ... "excellent exposures of which" ... "can be traced along the left wall of the Terek River valley between the villages of Kazbegi and Pansheti" (p. 18). In this section, which can be regarded as the stratotype, the Kazbek Formation can be easily subdivided in two subformations, like everywhere in the region.

The lower Kazbek Subformation (terminal lower Toarcian) reveals an irregular alternation of members (10–60 m) of massive sandstone beds 0.5–2.5 m thick, which are separated by minor interlayers of banded silty shale, and other members (15–90 m) of thin-banded silty shale, in which sandstone interlayers may occur as well. Sandstones are thin-banded, fine-grained, micaceous, massive at the stratum base and displaying horizontal to convolute bedding higher. In some strata, gradation texture and flysch hieroglyphs have been observed at the base. In general, such members often have a flyschoid habit. Proportions between sandy and clayey–silty members are variable. In the Fiagdon River heads, the clayey–silty members are of subordinate significance, they enclose frequent sandy interlayers, and their thickness does not exceed 10–30 m, whereas sandy members constitute to 60–70% of the whole section. In northern exposures of the Khevsurskaya Aragvi River valley, on the contrary, the sandy members are of a minor importance; they are more rare, while distinctly banded clayey–silty members increase in thickness to 60–80 and even 120 m, and exhibit a fine alternation of clayey–silty and sandy–silty laminae. In the habit, the lower Kazbek Subformation is similar at this site to the lower Lars Subformation of the Pseashkho zone. In the southernmost exposures along the Khevsurskaya Aragvi River, sandstone beds constituting 50–80% of the lower Kazbek Subformation rhythmically alternate with subordinate beds of banded silty shale. These sandstones of distinctly graded texture exhibit the crisscross, wavy, or convolute bedding and large casts of erosion grooves at the base of strata. The whole sequence has a typical habit of the proximal flysch. The apparent thickness of the lower Kazbek Subformation does not exceed 500–800 m in separate tectonic slices. In the only complete section of the Fiagdon River heads, the subformation is 800 m thick.

The upper Kazbek Subformation (upper Toarcian) represents everywhere a sequence of clearly banded silty shale enclosing very frequent interlayers of sandy siltstone and sandstone. It is only in places that members (up to 40 m) with contiguous sandstone strata up to 0.5–1.5 m thick can be distinguished. In some areas (Mnaisidon River, Terkhen River, the right-bank area of Snostskali River), the subformation has a typical habit of rhythmical distal flysch, thin members of which consist of sandy siltstone grading into fine-grained sandstone (2–5 cm), and then into shale (8–15 cm), all alternating with individual but rather abundant sandstone strata (20–35 cm). The basal plane of strata is distinct, and their upper portions reveal wavy and cross lamination. In places, these flyschoid rhythms are thicker: shale intervals up to 30–50 cm thick and sandstone

interval up to 5–25 cm thick. In this case, typical of sediments are the convolute bedding in the upper bed parts and casts of erosion grooves at the base.

I examined the boundary between the upper Kazbek Subformation and the overlying Gudoshaur Formation in upper reaches of the Mnaisidon and Terkhen rivers, in the left bank area of Snostskali River, and along Gudanischala River (left tributary of Khevsurskaya Aragvi River in its upper reaches). In all localities it is conformable, and the same was noted by other researchers in head areas of the Ardon and Terek rivers and in the Assa–Khevsurskaya Aragvi watershed. The thickness of the upper Kazbek Subformation is 1025 in the only complete section at the right bank of Gudanischala River. The apparent thickness of individual tectonic slices varies from 320 (left bank area of the Snostskali River) to 1140 m (Terek River valley).

All the reported fossils have been found in the east, in Khevsuretia; they characterize deposits of the Kazbek Formation without a precise reference to subformations distinguished later on. The Toarcian belemnites *Mesoteuthis conoidea* Opp., *M.* cf. *tripartita* Schl., *M.* cf. *triscissa* Jan., and *Dactyloteuthis* sp. were found in the Chanchakhis–Tskali River valley (right source of the Argun River) near the Village of Khakhabo (Krestnikov, 1944). In the Argun River heads, P.I. Avalishvili found the early Toarcian ammonites *Dactylioceras* sp. and *Nautilus austracoides* Joung et Bird, the latter definitely characterizing the lower formation part. Ammonites *Denckmannia* sp., *Haugia* sp., *Grammoceras thouarsense* Orb., *Erycites* sp., *Dumortieria* cf. *bleicheri* Ben., and *Pleydellia* cf. *subcompta* Br., all found in upper reaches of the Khevsurskaya Aragvi River (Giorgobiani *et al.*, 1983), characterize all three upper Toarcian zones. Hence, the Toarcian age of the Kazbek Formation is beyond questions. The distinguished subformations are correlative with concurrent subdivisions in sections of the Pseashkho and Digoria–Osetia zones that verifies their ages and the total stratigraphic range of the Kazbek Formation (Fig. 2). The lower Kazbek Subformation is correlative, in lithology and stratigraphic position, with the lower Lars Subformation and Ksurt Formation, and accordingly corresponds to the upper part of the *H. falcifer* Zone and the *H. bifrons* Zone of the lower Toarcian. Exactly their interval yields the early Toarcian fossils mentioned above. The upper Kazbek Subformation is also reliably correlated with the upper Lars Subformation and the Dzhimara Formation and referred to the upper Toarcian. Judging from description of enclosing rocks, the upper Toarcian ammonites of Khevsuretia were recovered from that subformation.

The Gudoshaur Formation (an upper part of the *D. levesquei* Zone, the upper Toarcian–Aalenian) was distinguished by Rengarten as "the formation of shales" conformably overlying the Kazbek Formation and separated from the overlying Busarchili Formation and Upper Jurassic rocks by the Tib overthrust plane. Rengarten did not suggest the formation stratotype. I observed characteristic sections of the Gudoshaur Formation along the Mnaisidon River at the right bank of Terek River, in valleys of the Snostskali River and its tributaries, and along the Khevsurskaya Aragvi River. Everywhere, the formation represents a uniform sequence of black shales with sporadic interlayers of siderite concretions and concretionary conglomerate. Rare interlayers of large bun-shaped siderite concretions are observable almost in all sections. In places, there are lenticular interlayers of brown limonitized siderite. Sandstone interlayers are rare as a rule. Only in the extreme south, in the Zakka River basin, the amount of sandstone and siltstone interlayers slightly increases. According to my observations, the apparent thickness of the formation is 900–1350 m. The total thickness of the Gudoshaur (Gudan) Formation in the Khevsurskaya Aragvi River basin is as great as 1500 m (Giorgobiani *et al.*, 1983).

Krestnikov found (1944) *Pleydellia mactra* Dum. from the lower part of the Gudoshaur Formation in exposures of the Pshavskaya Aragvi River (left tributary of Khevsurskaya Aragvi River), and *Leioceras opalinum* Rein. is known from the Mt. Arkhoti site in the Assa–Khevsurskaya Aragvi watershed area (Giorgobiani *et al.*, 1983). These forms characteristic of the uppermost *D. levesquei* Zone point to the upper Toarcian–lower Aalenian range of the formation that can be correlated, according to lithology and stratigraphic position, with the Salgi Formation of the Digoria–Osetia zone, as Panov and Shevchenko (1964) suggested. Present-day data imply however that the Gudoshaur Formation likely spans the entire Aalenian Stage. According to Enna, who studied this subdivision along the Zakka River in the Ardon River heads, it is conformably overlain by the Bursachili Formation, which is not older than the Bajocian. The same relationship is established in exposures of the Khevsurskaya Aragvi River (Giorgobiani *et al.*, 1983). In this case, the upper part of the Gudoshaur Formation is correlative with the Koirakh and Fortaukh formations (Fig. 2). As an argument in favor, I can mention the upper member of the Gudoshaur Formation exposed along the Terkhen River. That member (up to 200 m thick) of shales alternating with fine-grained thin-banded sandstone and sandy siltstone could be an analog of the Koirakh Formation. Similar member of shale and sandstone beds alternating in the upper part of the Gudoshaur Formation are also distinguishable in its thickest sections exposed along the Snostskali River and at the right bank of the Terek River.

In the Assa–Khevsurskaya Aragvi watershed, Kipiani distinguished a sequence (500–650 m) of shales with sandstone interlayers above the Gudoshaurskaya Formation in the core of the Arkhot syncline. Sandstone beds with flysch hieroglyphs and with mytiloids *Mytiloceramus cinctus* (Goldf.) and *M. quenstedti* Pcel. dominate in the lower part of the sequence, whereas shales with siderite concretions prevail in its upper por-

tion. Kipiani termed this sequence as the Bursachili Formation. However, it likely represents the upper part of the Gudoshaur (Gudan) Formation, as Giorgobiani *et al.* (1983) believe. In that instance, the lower part of the sequence would be an analogue of the Koirakh Formation, while the upper one could be correlated with the Fortaukh Formation. If this is correct, the Gudoshaur Formation should span both the upper Aalenian and the lowermost Bajocian.

The Bursachili Formation (Bajocian to Bathonian ?) was distinguished by Rengarten south of the Tib overthrust within the flysch synclinorium, where, according to his data, it is exposed in anticline cores. At present, deposits of the Bursachili Formation are also established northward of the Tib overthrust, within a belt extending from the Zakka River in the west to the Khevsurskaya Aragvi River in the east. In all the sections studied, the formation constitutes tectonic sheets separated by a thrust plane from deposits of the Gudoshaur or Kazbek formations in the north and by the Tib overthrust from the Upper Jurassic flysch in the south. Conformable stratigraphic relationships between the Bursachili and Gudoshaur formations have been observed only in some places of the Zakka and Khevsurskaya Aragvi river basins.

Rengarten did not suggest the stratotype for the Bursachili Formation. I observed its typical, well-exposed sections in lower reaches of the Mnaisidon River, at the left bank of the Snostskali River, and along the Khevsurskaya Aragvi River. Enna described the formation section in the Zakka River basin. In all sections, the formation exhibits frequent, in places flyschoid alternation of shale and brownish–gray calcareous sandstone beds with distinct flysch hieroglyphs at the base. According to relative proportions of particular rocks, the Bursachili Formation is divided in two subformations. *The lower Bursachili Subformation* is a sequence of shales with horizons of siderite concretions, with frequent sandstone (0.1–0.3 m) and thin-banded sandy siltstone interlayers. Sandstones are gray to brown, fine-grained, calcareous, displaying fine horizontal bedding that gives place upward to crisscross wavy and graded lamination. Flysch hieroglyphs or casts of erosion grooves are well expressed at the base of nearly all sandstone beds, which constitute up to 30% of the total thickness. In separate tectonic slices, the apparent thickness of the unit is 470–530 m. *The upper Bursachili Subformation* is lacking siderite concretions, and its sandstone beds of a greater thickness (1–2 m) constitute up to 50% of some members. In places, sandstone beds enclose pebbles of shales. In the left bank area of the Snostskali River, the apparent thickness of the subformation is not greater than 400 m. According to my observations along Khevsurskaya Aragvi River, the entire Bursachili Formation is up to 1100 m thick.

Stephanoceras sp., found in the Zakka River basin by I.G. Kulaev, indicates that lower Bajocian beds are present in the Bursachili Formation. The formation is similar in lithology and stratigraphic position to the Bzych Formation of the northwestern Caucasus and to the Zurunchai and Gdym formations of the same tectonic zone in the southeastern Caucasus, where they are dated back to the Bajocian–Bathonian (Panov and Gushchin, 1987) and are conformably overlain by Callovian deposits. This suggests the upper Bajocian–Bathonian range of the Bursachili Formation.

The relationship between the Bursachili Formation and Callovian–Upper Jurassic flysch of the Chiauri synclinorium is still unclear, since the contact between them is tectonic (the Tib thrust fault). In the left bank area of Terek River opposite the Village of Kobi, there were found breccia-conglomerate lenses composed of rock fragments of the Bursachili Formation near the thrust fault zone (Panov and Shevchenko, 1964). These rocks, formerly accepted for a basal flysch horizon, most likely represent the intraformational conglomerate, because among rock fragments there are also marls of the flysch sequence. This does not rule out the stratigraphic unconformity (a hiatus) between the Bursachili Formation and flysch sequence, though the angular unconformity is certainly absent here. These and those rocks are constituents of a single imbricated–nappe structure of the southern flank zone.

CONCLUSION

Formations and subformations distinguished in the Liassic–Aalenian sequences of the northern, axial, and southern zones of the Greater Caucasus are well correlative at the corresponding stratigraphic levels, which are used to establish the formation ages in the axial part and southern slope and to subdivide them into substages and zones. In particular, the Early Jurassic age of the Kistinka and Tsiklauri formations is reliably substantiated for the first time despite the former opinions that they are of the Triassic or even Paleozoic are. This has been done based on uniform sedimentation cycles characteristic of a basin that existed on the territory of the Greater Caucasus during the Liassic–Aalenian period. For the Bajocian–Bathonian interval, the precise correlation between formations of the northern and southern zones is impossible, and the Bursachili Formation of southern areas is dated conventionally to a certain extent, because a system of central uplifts, which developed since the initial Bajocian in the axial part of the Greater Caucasus, separated two parts of the former single basin.

The precise age determination for all the formations distinguished by Rengarten opens a possibility to regard the Lower–Middle Jurassic sequence exposed along the Georgian Military Road as the reference one for the axial and southern zones of the Greater Caucasus. Correlating the formations with those of the Upper Triassic–Middle Jurassic sequence in the Mountainous Crimea (Panov, 1997a), we can verify the relationship between Jurassic and pre-Jurassic deposits in the study

region, elucidate stratigraphy and ages of Lower Jurassic subdivisions, clarify relationships between the Liassic–Aalenian and Bajocian–Bathonian intervals, and solve many other problems.

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Stratigraphy and Geological Correlation, Vol. 11, No. 1, 2003, pp. 74–90. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 11, No. 1, 2003, pp. 78–95.
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Comparative Analysis of the Eocene–Miocene Micro- and Macrofloras of the Eastern Sikhote Alin'

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Received June 6, 2001

Abstract—Comparative characteristic of spore–pollen assemblages from the Son'e (late Eocene), Bui, Svetlovodnaya (late Eocene–early Oligocene), Amgu (late Oligocene–early Miocene), Dembi, Velikaya Kema (late early Miocene–earliest middle Miocene), and Botchi (latest middle Miocene–late Miocene) floral localities in the Eastern Sikhote Alin' are presented. It is shown that assemblages of micro- and macroflora are substantially similar at the generic and family levels. They usually include the following genera in common: *Ginkgo, Pinus, Picea, Tsuga, Sciadopitys, Thuja, Alnus, Betula, Carpinus, Ostrya, Carya, Fagus, Quercus, Acer,* and *Rhus.* Differences between assemblages in composition and abundance of taxa are determined by biological peculiarities of plants, fossilization mode of micro- and macrofloras, geographical position of particular areas, conditions of lithogenesis, and genesis of assemblages; in addition, morphology of some spore and pollen species is insufficiently known.

Key words: microflora, macroflora, integral floral analysis.

INTRODUCTION

Biostratigraphic subdivision of Cenozoic continental sequences of the Far East is based exclusively on paleobotanic data (leaf impressions and remains of plant fruits, seeds, spores, and pollen). The analysis of macroflora reveals composition of phytocoenoses formerly developed in the accumulation area of flora-bearing sediments, i.e., allows recognition of past local, usually intrazonal plant assemblages. Spores and pollen can be transported by wind over a significant distance, and palynological assemblages reflect therefore the composition of vegetation that grew in areas rather remote from macroflora burials and thus enable interpretation of the zonal vegetation type. Of course, some allochthonous elements occur in macroflora assemblages as well, although distance of transportation of macroflora remains is significantly lesser than that of pollen (up to several hundreds of kilometers for some pollen genera).

It is well known that plant species and genera differ in productivity of spores and pollen. The long period of investigations showed that composition of the "pollen rain" differs from that of spore–pollen spectra, i.e., the latter do not reflect adequately the diversity and composition of vegetation (Zaklinskaya and Grichuk, 1948; Kabailene, 1969; Kondratene, 1976; and others), and the difference degree was estimated. Composition of spore–pollen spectra in recent sediments is primarily influenced by biological factors, such as the plant pollination mode, capability of transportation by water and wind, morphology of pollen grains, their specific weight, and chemical composition of exine, and also by physicochemical conditions in a particular area. It is needless to say that subfossil and fossil spectra differ in composition as well because of progressing destruction of palynological remains in sediments. To a significant extent, this is a consequence of exine preservation in different species depending on its chemical composition, lithology, and sedimentation rates. As a result, the composition of fossil spore–pollen spectra differ usually from that of fossil macroflora. Preservation of macrofloras depends on habitat environments. Fossil floral assemblages characterize usually hydrophilic and mesophilic vegetation of lowlands (marshes, lakes, and river coasts) and low mountains. In addition, macroflora assemblages are dominated, as a rule, by tree leaves, whereas herbaceous remains are commonly decomposed in situ and not preserved. Thus, the simultaneous analysis of plant impressions and palynological fossils is of prime importance in paleobotanic investigations. Combination of these two methods leads to the more adequate reconstruction of fossil flora composition that is important for more accurate dating, stratigraphic subdivision of sequences, and climatic or paleogeographic interpretations.

The combination of two paleobotanic methods was previously used to study Miocene sediments along the Botchi River in the Far East (Akhmet'ev, 1973a), Miocene sequences of the Mt. Mamontovaya area at the Aldan River (V.P. Nikitin, I.A. Il'inskaya, and A.F. Fradkina in *Miotsen...*, 1977), Neogene sequences of Northeastern Asia (Fradkina, 1983), and Jurassic deposits of eastern Central Asia (Genkina and Dubrovskaya, 1979, 1984).

In this work, I analyzed composition of spores and pollen macerated from samples with plant impressions, which nave been collected from seven floral localities of the Pribrezhnyi basaltic belt in the Eastern Sikhote Alin'. Floras from these localities range in age from the late Eocene to late Miocene. The defined spore–pollen spectra (SPS) are thoroughly compared with assemblages of macroflora, and probable factors responsible for similarity and difference in their qualitative and quantitative composition at the generic level are analyzed.

HISTORICAL REVIEW

The Son'e, Bui, and Dembi flora localities along the coast of the Tatar Strait between the Son'e and Dembi bays and those of the Svetlovodnaya, Amgu, and Botchi rivers, and near the Velikaya Kema Settlement were chosen for the study (Fig. 1). Plant impressions from these localities were examined earlier (Akhmet'ev, 1965, 1973a, 1973b, 1974, 1988; Akhmet'ev and Shevyreva, 1989; Akhmet'ev and Vikulin, 1995; Akhmet'ev and Manchester, 2000). Data by R.S. Klimova for the Amgu and Velikaya Kema localities (see in Rybalko *et al.*,) were also used in the analysis. In my earlier publications dedicated to the study of spores and pollen from the Amgu, Son'e, Bui, Dembi, and Velikaya Kema localities (Lopatina 1999, 2001a, 2001b), I estimated, using both methods, the age of the sequences in question and correlated the spore–pollen complexes of the regional. Ages of flora-bearing deposits, which have been inferred from spore–pollen complexes and plant impressions, coincide and correspond to late Eocene at the Son'e locality, late Eocene–early Oligocene at the Bui and Svetlovodnaya localities, late Oligocene–early Miocene at the Amgu locality, late early–initial middle Miocene at the Dembi and Velikaya Kema localities, and terminal middle–late Miocene at the Botchi locality.

It should be noted that vegetation of the territory under consideration developed in the Late Paleogene–Miocene under a strong influence of an intense volcanic activity. Most flora-bearing localities of the Pribrezhnyi volcanic belt are confined to sediments accumulated in oxbow and small dammed lakes. Oleinikov (1977) divides the lakes in two types. The first type includes lakes of several dozens square kilometers in size, which were located in large grabens (Amgu locality). Flora-bearing sediments of the Son'e, Bui, Dembi, Svetlovodnaya, and Velikaya Kema localities accumulated in lakes of the second type. These short-lived lakes originated in response to damming of small rivers and creeks by lava flows or landslides and were evidently as large as several square kilometers. Flora-bearing lacustrine sediments of the Botchi section accumulated beyond the influence zone of active volcanoes. All examined palynological preparations contain *Botryococcous* and *Ovoidites* sp., the last of which is an analog of contemporary Zignemaceae algae of the *Spirogyra* genus characteristic of the freshwater sedimentation setting.

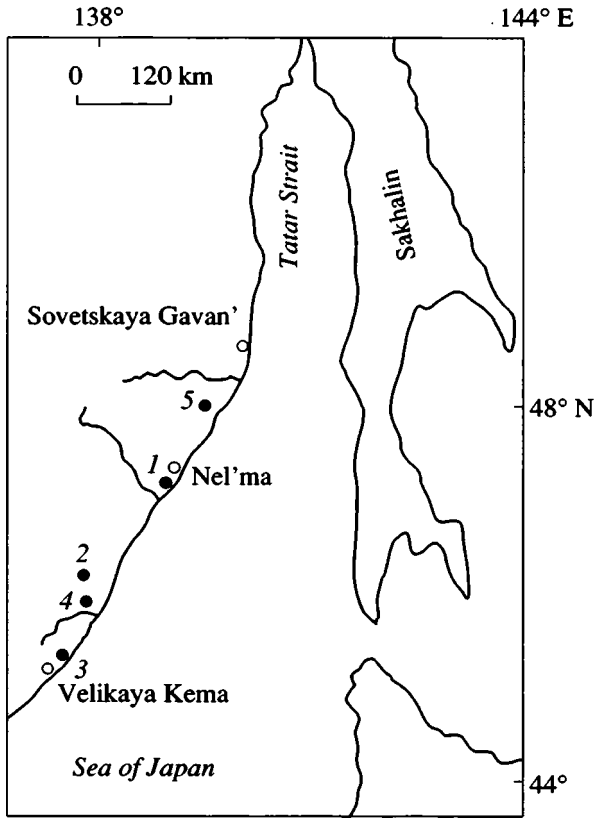


Fig. 1. Localities of the studied (*1*) Son'e, Bui, Dembi, (*2*) Svetlovodnaya, (*3*) Velikaya Kema, (*4*) Amgu, and (*5*) Botchi sections.

MATERIALS AND METHODS

In total, 55 samples with plant impressions from collection by Akhmet'ev were used for the palynological analysis. They characterize seven localities of the Eastern Sikhote Alin'. These are the Amgu, Botchi (10 samples from each), Bui (8 samples), Son'e, Dembi, Velikaya Kema (7 samples from each), and Svetlovodnaya (6 samples) localities. Samples were treated at the Laboratory of Paleofloristics, Geological Institute RAS, and at the Laboratory of Biostratigraphy, Institute of the Lithosphere of Marginal Seas RAS, using the Grichuk maceration method and Erdtman acetolysis procedure. When examining preparations, I paid special attention to abundance and preservation of palynomorphs. In some samples from the Svetlovodnaya (one), Amgu (two), and Botchi (two) localities, the content of palynological remains did not exceed ten grains per a preparation, and they were excluded from the subsequent analysis.

Numbers of genera and families in the micro- and macroflora assemblages from examined sections of the Eastern Sikhote Alin'

Number of genera	Son'e	Bui	Svetlovodnaya	Amgu	Dembi	Velikaya Kema	Botchi
Total quantity	72	83	68	61	70	72	69
Number of genera identified in spore–pollen spectra	60	76	62	46	41	56	46
Number of genera identified amid plant impressions	33	21	17	40	50	38	43
Number of genera identified in spore–pollen spectra and amid plant impressions	22	15	11	22	21	23	17
Number of families							
Total quantity	46	52	43	34	39	45	42
Number of families identified in spore–pollen spectra	36	47	38	25	26	35	29
Number of families identified amid plant impressions	21	16	15	24	25	24	27
Number of families identified both in spore–pollen spectra and amid plant impressions	14	11	9	15	12	12	11

RESULTS AND DISCUSSION

Comparison of micro- and macroflora composition at the generic level (table) shows that the number of plant genera determined by palynological method is usually higher than that of macroflora remains. This can be explained by above-mentioned differences in the accumulation mode and preservation of palynological and macroflora remains. The number of families established in palynological analysis is equal to ore slightly exceeds that in macroflora assemblages. The number of genera and families established in the examined sections, using simultaneously two methods, ranges from 11 to 23 and from 9 to 15, respectively.

It should be noted that macrofossils of some genera (e.g., of *Acer* and *Carpinus*) have distinct morphological features allowing their identification at the species level, whereas their pollen can be determined at the generic level only because of a poor preservation state. On the other hand, morphological peculiarities of some pollen taxa (e.g., *Tsuga* pollen) are appropriate for the species identification among corresponding palynomorphs (for instance, eight species at the Botchi locality). To the contrary, seeds and cones of this plant can be identified at the generic level only. The genetic relation between fossil palynomorphs and vegetative parts of plants can be established only in the case, when pollen is extracted from fertile organs. Investigations of this kind were carried out by Krasilov (1966). Presumable identification of micro- and macrofossils can be sometimes performed analytically. Par example, Betulaceae pollen bearing simultaneously features of *Carpinus* and *Corylus*, as it was established in the SPS of the Bui locality by N.I. Zaporozhets and later by myself, can be referred to the species *Paleocarpinus sikhotealinensis* that was described by Akhmet'ev and Manchester (2000) from the same locality as having features characteristic of both the hornbeam and hazel.

COMPOSITION AND ABUNDANCE OF MICRO- AND MACROFLORAS

Let us consider the qualitative and quantitative composition of palynological and plant assemblages from every locality (Figs. 2–8). The Son'e SPS is remarkable for a high content of spores with smooth exine (over 20% in some spectra), which belong mainly to leguminous Polypodiaceae, although impressions of corresponding sporogenous plants are absent (Fig. 2). Gymnosperms in the SPS are mainly represented by pollen of *Pinus* and *Taxodium* accompanied by a lower quantity of pollen identified as Taxodiaceae or Cupressaceae because of a poor preservation state. Preponderant among impressions of gymnosperms are those of Taxodiaceae (*Cryptomeria*), Cupressaceae (*Thuja*), and Pinaceae (*Pinus*). The angiosperm spore–pollen group is dominated by Betulaceae (*Alnus, Betula, Corylus, Carpinus*) and *Quercus* pollen. *Fagus, Quercus, Cocculus, Trochodendron, Cercidiphyllum*, and *Platanus* prevail amid plant impressions, whereas the content of *Trochodendron* and *Cercidiphyllum* in the SPS does not exceed 0.5%, and pollen of *Cocculus* and *Platanus* is missing.

In the flora from the Bui locality, sporogenous plants are absent among impressions (Fig. 3). Gymnosperms are represented by noticeable amount of *Pinus, Taxodium*, and Taxodiaceae or Cupressaceae specimens in the SPS and only by *Thuja* impressions in macroflora. Angiospermous macroflora is dominated by *Tilia, Macaranga, Acer*, and *Phragmites* whose pollen is either absent in the SPS or occurs in an insignificant quantity. Missing also is pollen of genera *Populus, Rosa, Eucommia, Machilus*, and *Alangium* represented in collection of plant impressions. A noticeable proportion in the macroflora belongs to *Alnus* and *Quercus* whose content is slightly higher than that of their pollen in the palynological assemblage, whereas the *Carya* content is, to the contrary, significantly lower.

Similarly to the Son'e and Bui macroflora assemblages, collection of plant impressions from the Svetlovodnaya River locality lacks sporogenous plant taxa (Fig. 4). The SPS is dominated here by Taxodiaceae pollen accompanied by subordinate share of Pinaceae pollen. Approximately half of all impressions characterize the *Metasequoia* genus represented by single grains in the SPS. *Ginkgo* is present in a significant proportion. Dominant angiosperms in macroflora collection are *Zelkova* and *Cocculus*, while pollen of the former occurs as single grains and the latter taxon is completely missing in the SPS. Angiosperms in the SPS represent mainly the genus *Carya* missing from impressions, and associated Betulaceae (*Carpinus*) and Myricaceae (*Comptonia*). *Carpinus* is characterized by approximately equal proportions in both the SPS and macroflora assemblage.

Preservation of spores and pollen in macerates from the Amgu section is poor, frequently hampering their identification at the generic level. Among sporogenous plants rare in both assemblages, *Osmunda* represents a taxon in common (Fig. 5), and the SPS and macroflora are characterized by prevalence of gymnosperms. Preponderant in the SPS is pollen of *Pinus*, *Taxodium* and poorly identifiable Taxodiaceae gen. indet. and Taxodiaceae or Cupressaceae. Analysis of macroflora revealed prevalence of Taxodiaceae (mainly of *Metasequoia* and *Glyptostrobus*) over subordinate Pinaceae. Angiosperms in the SPS are largely represented by *Acer, Ulmus*, and also tricolporate pollen of indeterminable taxonomic position. Preponderant amid plant impressions are *Alnus* (below 0.5% in the SPS) and different representatives of frutescent genera *Spirea, Sorbus, Syringa, Lonicera, Viburnum*, and others whose pollen is missing. P.S. Klimova (in Rybalko *et al.*, 1980) determined in this locality the plant impressions similar in general to those described by Akhmet'ev. *Metasequoia* is very abundant among them, whereas *Glyptostrobus* that prevails in Akhmet'ev's collection has not been established. Klimova also mentioned genera *Sequoia, Carya*, and *Corylus* whose pollen is identified also in the spore–pollen spectra (Lopatina, 1999).

The Dembi locality yielded the SPS of two types. Spectra of the first type are characterized by prevalence of *Fagus* pollen amid angiosperms representing to 30% of palynomorphs. Pollen of gymnosperms characterizes mainly Pinaceae (about 20%) with a noticeable share of Taxodiaceae (*Taxodium*). The peculiar feature of palynological spectra of the second type is preponderance of gymnosperm pollen of the family Pinaceae (*Pinus, Picea, Abies*). *Quercus* and *Fagopsis* are dominant among angiosperm and *Metasequoia* among gymnosperm plant impressions. Impressions of Pinaceae

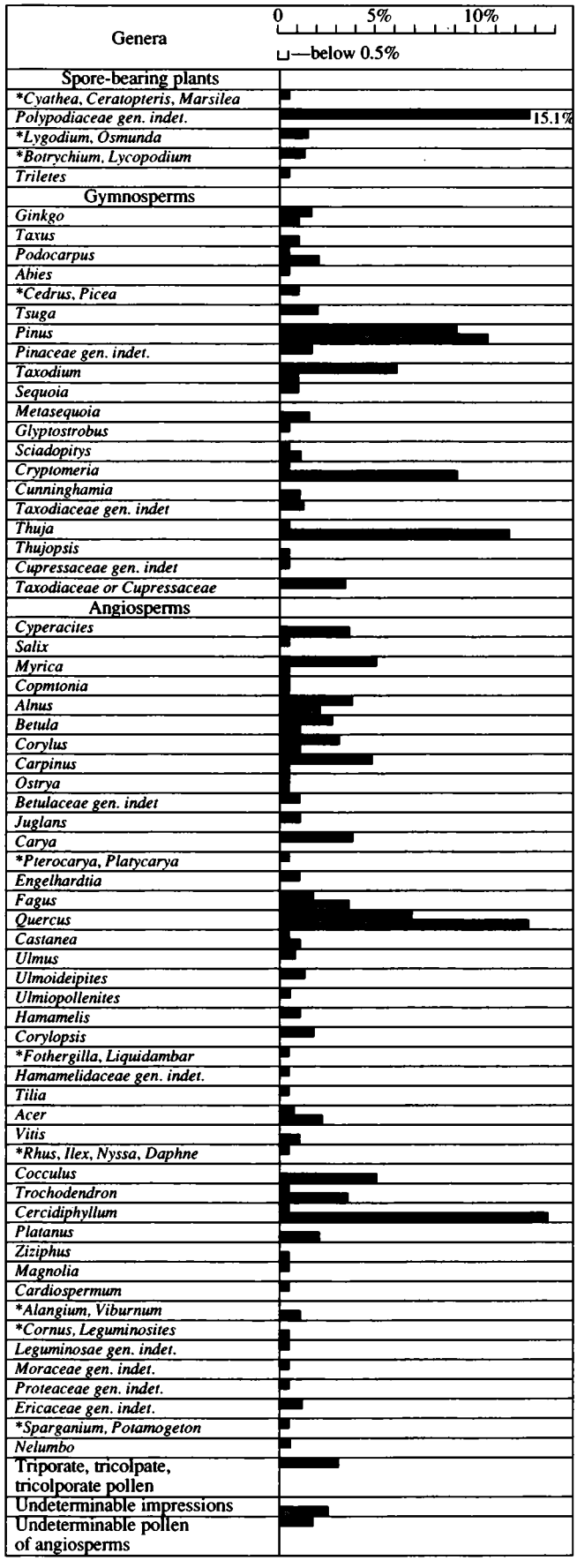


Fig. 2. Diagram illustrating variations in percentages of palynomorph (black) and macroflora (gray) taxa at the Son'e locality (genera of equal abundance are marked by asterisk in Figs. 2–5, 7, 8).

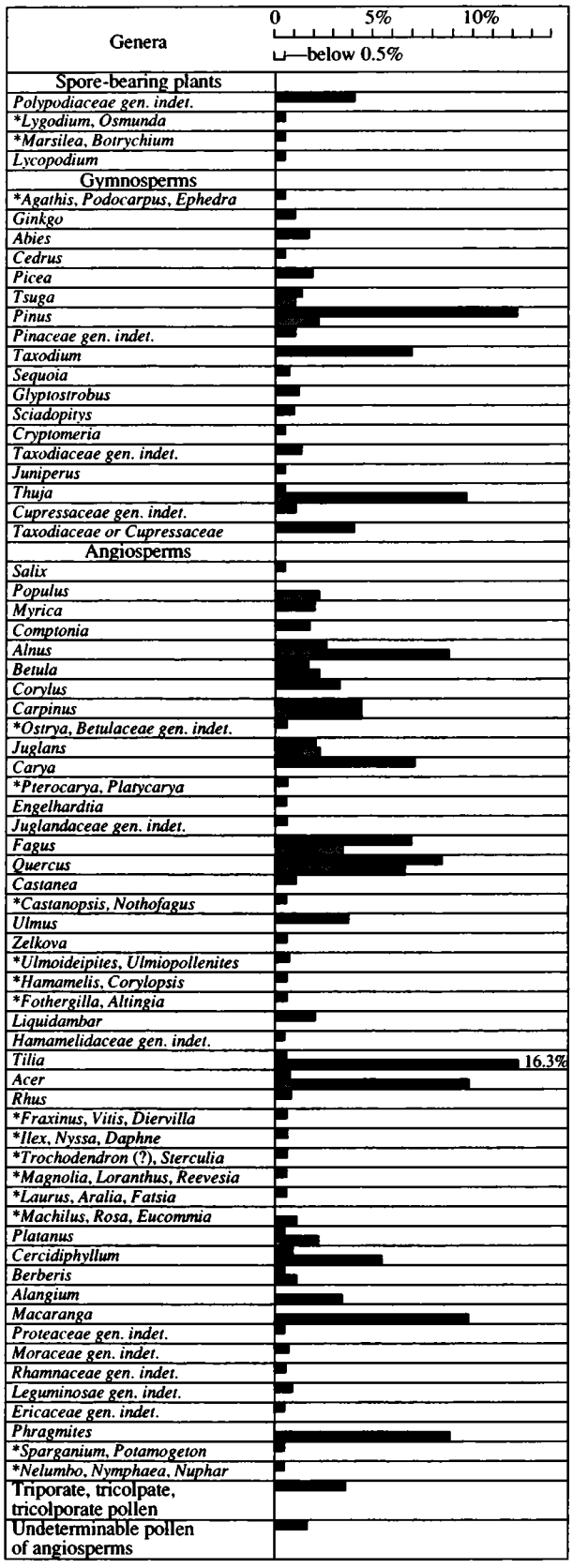


Fig. 3. Diagram illustrating variations in percentages of palynomorph (black) and macroflora (gray) taxa at the Bui locality (symbols as in Fig. 2).

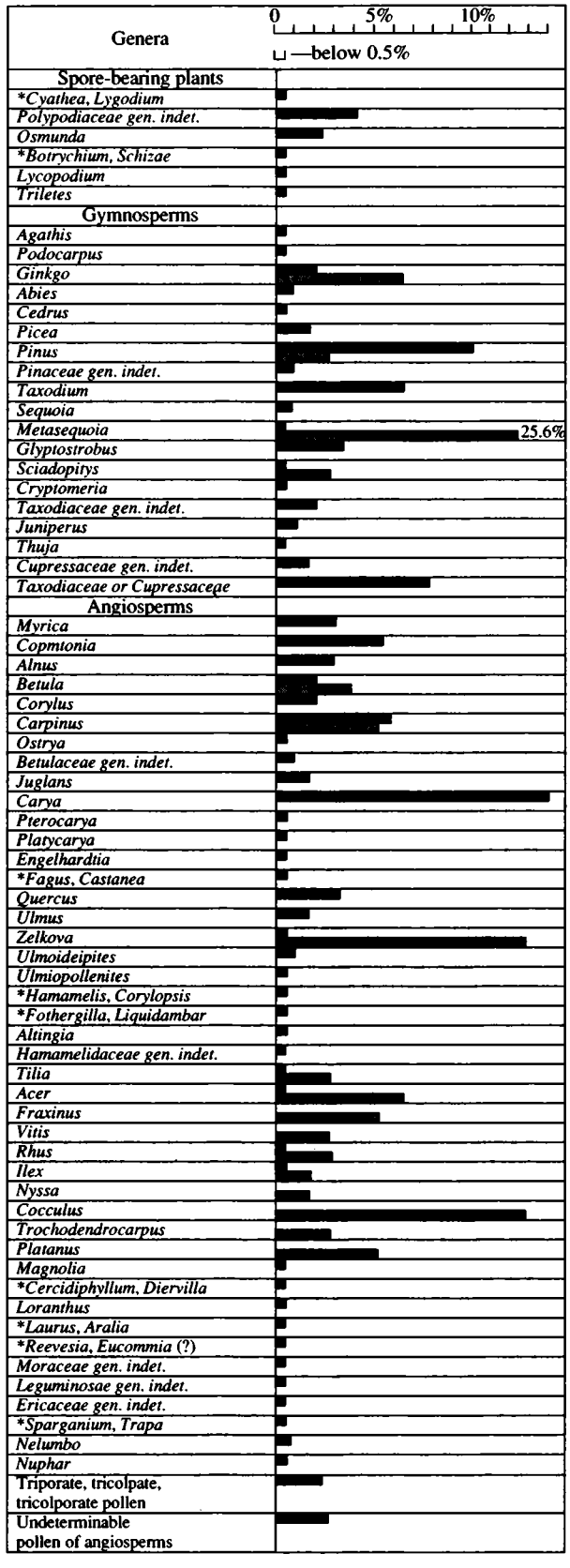


Fig. 4. Diagram illustrating variations in percentages of palynomorph (black) and macroflora (gray) taxa at the Svetlovodnaya locality (symbols as in Fig. 2).

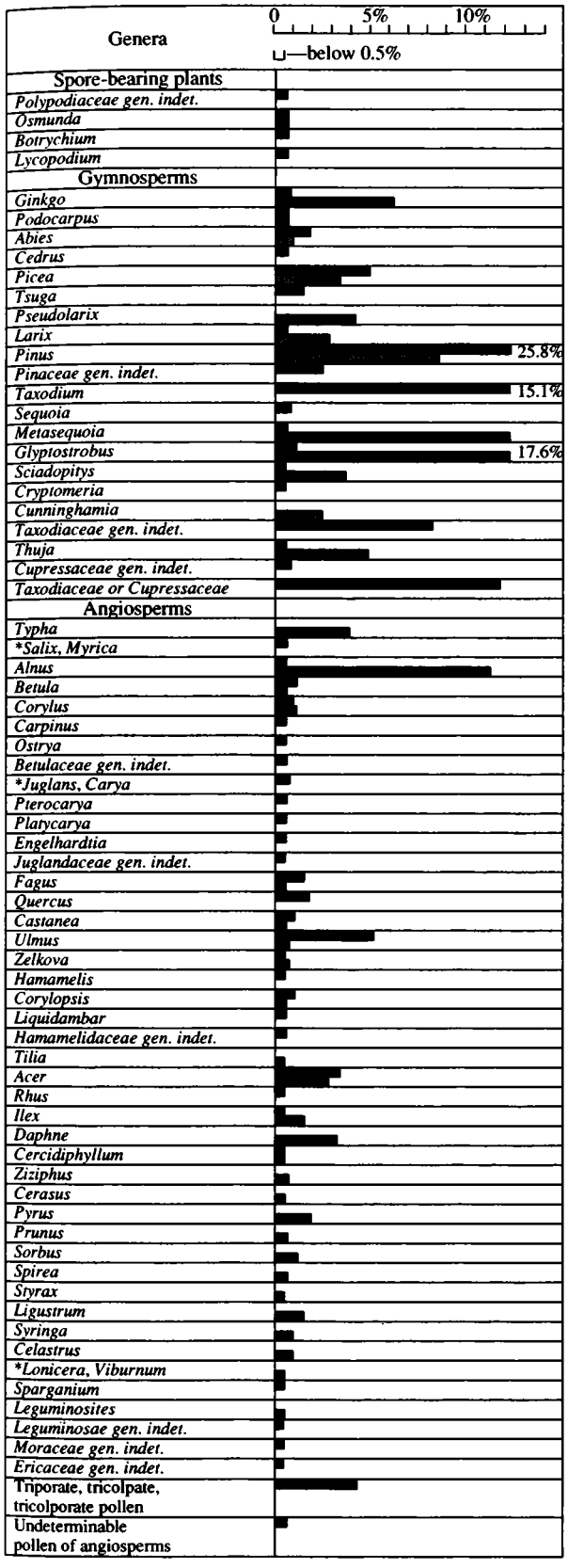


Fig. 5. Diagram illustrating variations in percentages of palynomorph (black) and macroflora (gray) taxa at the Amgu locality (symbols as in Fig. 2).

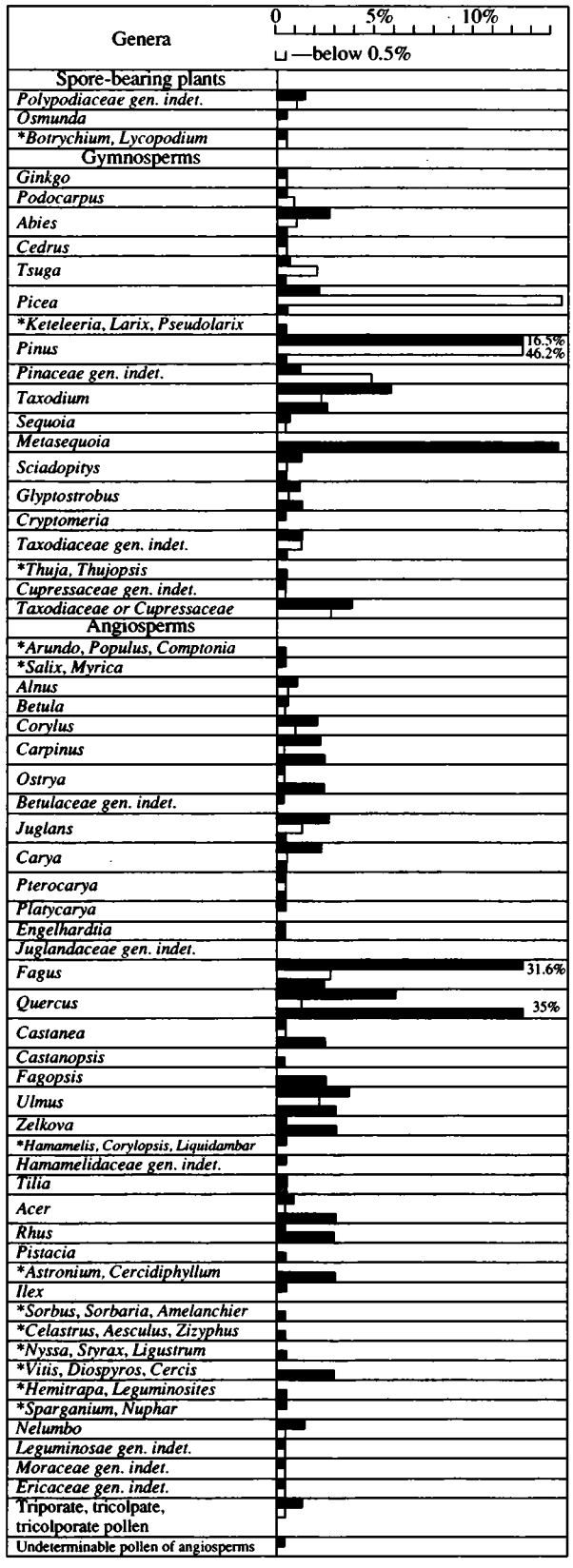


Fig. 6. Diagram illustrating variations in percentages of palynomorph taxa in the spore–pollen spectra of the first (black) and second (white) types in comparison with those of macroflora genera (gray) at the Dembi locality.

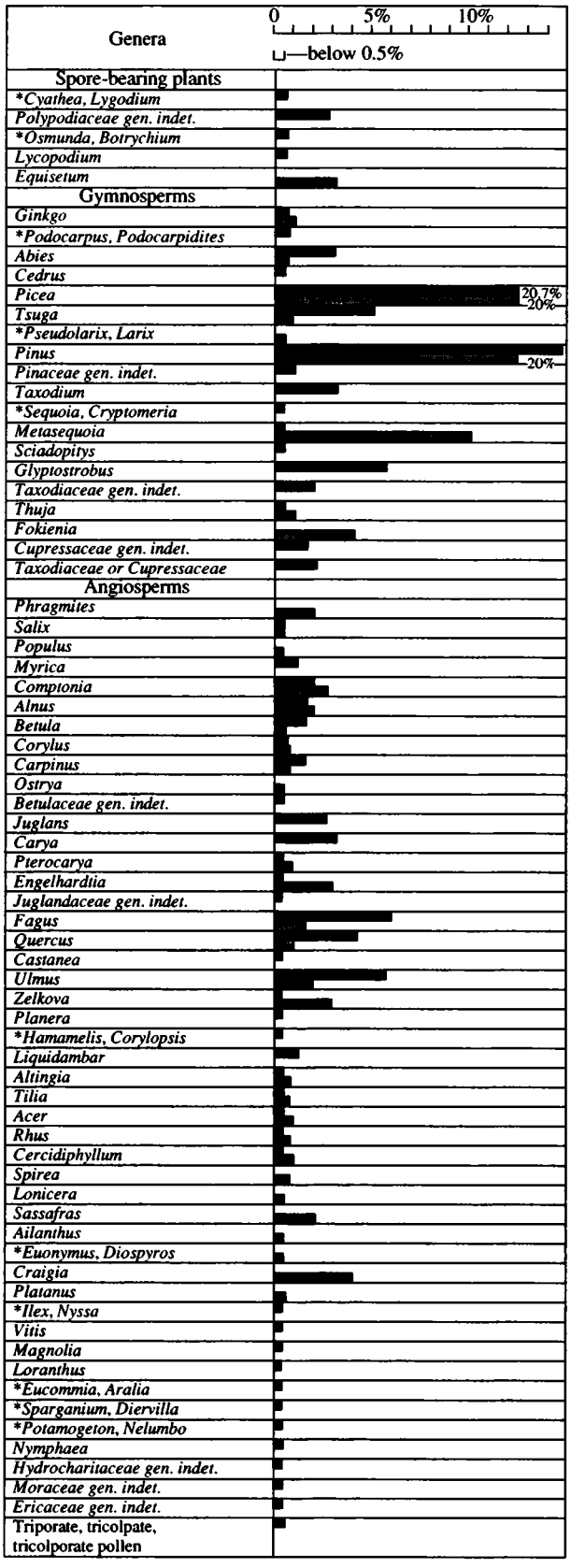


Fig. 7. Diagram illustrating variations in percentages of palynomorph (black) and macroflora (gray) taxa at the Velikaya Kema locality (symbols as in Fig. 2).

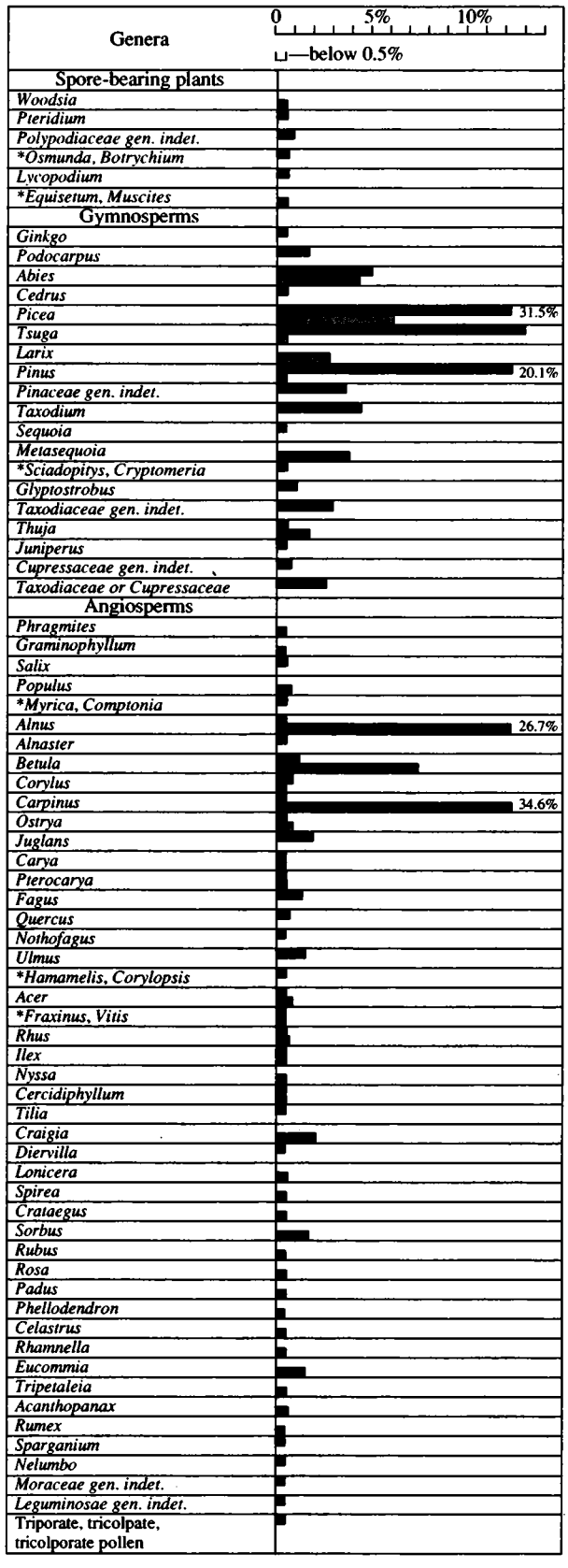


Fig. 8. Diagram illustrating variations in percentages of palynomorph (black) and macroflora (gray) taxa at the Botchy locality (symbols as in Fig. 2).

are rare, although they are rather diverse at the generic level. As is seen in the table, the number of genera determined in macroflora exceeds significantly that identified by the palynological method. *Pistacia, Astronium, Cercidiphyllum, Vitis, Nyssa, Diospyros*, and genera of the family Rosaceae are established only in the macroflora collection.

The SPS from the Velikaya Kema locality is dominated by pollen of *Picea* and *Pinus*, and both genera prevail, with comparable proportions, in macroflora as well (Fig. 7). *Tsuga, Taxodium*, and *Glyptostrobus* are of a high abundance in the SPS, but rare (*Tsuga*) or missing (*Taxodium, Glyptostrobus*) amid impressions. The content of *Metasequoia* pollen is about 0.5%, while this genus is third in abundance, after *Pinus* and *Picea*, among impressions. The number of angiosperm genera determined by both methods is quite comparable for the Betulaceae, Juglandaceae, Fagaceae, and Ulmaceae families. Percentages of *Alnus, Betula, Carpinus, Engelhardtia*, and *Zelkova* pollen are slightly lower in comparison with these genera abundance in macroflora collection. Simultaneously, the opposite relationships are established for *Fagus, Quercus*, and *Ulmus*. Genera *Corylus, Pterocarya, Acer, Rhus*, and *Cercidiphyllum* are characterized by an insignificant (usually below 1%) approximately equal contents in the spore–pollen and macroflora assemblages. Klimova identified single impressions of *Glyptostrobus, Cryptomeria, Potamogeton, Pterocarya, Carya, Castanopsis, Cercidiphyllum*, and *Vitis* (in Rybalko *et al.*, 1980), which occur also in the palynological spectra. Present in them is also *Juglans*, although the proportion of its impression in Klimova's collection is slightly higher (12).

The content of some sporogenous genera in palynological and macroflora assemblages from the Botchi locality is usually below 0.5% (Fig. 8). In the Polypodiaceae macroflora, there is established the genus *Woodsia*, and spores are identified as Polypodiaceae gen. indet. The SPS is dominated by gymnosperm pollen (about 90%). Prevalent are Pinaceae, namely *Picea, Pinus, Tsuga*, and *Abies* (in the order of decrease). In the subordinate pollen group of Taxodiaceae, the dominant *Taxodium* pollen is associated with specimens identified as Taxodiaceae gen. indet. and Taxodiaceae or Cupressaceae. The share of angiosperms pollen is insignificant (on the average about 10%); noticeable pollen contents are also characteristic of the Betulaceae and *Fagus, Quercus*, and *Ulmus* genera. In contrast, angiosperms, mainly Betulaceae (*Alnus, Carpinus*) dominate amid plant impressions. The share of gymnosperms in macroflora collection is about 18%, and impressions mostly represent *Picea, Abies, Metasequoia*, and *Larix* genera.

Most characteristic and well-preserved spores and pollen, which have been macerated from samples collected in seven macroflora localities, are illustrated in Plates 1–3.

METHODOLOGICAL ASPECTS OF COMPARISON BETWEEN SPORE–POLLEN AND MACROFLORA ASSEMBLAGES

The analysis of spore–pollen and macroflora assemblages from seven floral localities elucidates some general regularities in distribution of fossil plant remains of several families and genera. Pinaceae (*Pinus, Picea*), Taxodiaceae (*Taxodium, Glyptostrobus, Sciadopitys*), Cupressaceae (*Thuja*), Betulaceae (*Carpinus, Ostrya*), Juglandaceae (*Juglans, Carya*), Fagaceae (*Fagus, Quercus*), Ulmaceae (*Ulmus, Zelkova*), Cercidiphyllaceae (*Cercidiphyllum*), Anacardiaceae (*Rhus*), and Aquifoliaceae (*Ilex*) have been identified by both investigation methods. Genera of aqueous plants (*Sparangium, Potamogeton, Nelumbo*) as well as genera *Magnolia, Myrtus, Loranthus, Aralia, Fatsia, Liquidambar*, and others are identified using the palynological method alone, while *Rosa, Sorbus, Spirea, Celastrus, Lonicera, Viburnum, Styrax, Populus*, and *Diospyros* are determined amid plant impressions only.

Impressions of sporogenous plants have not been found in the Son'e, Bui, Dembi, and Svetlovodnaya floras, although spores are present in all examined spore–pollen assemblages. In some SPS from the Son'e locality, the share of spores exceeds 20%. Sporogenous plants occur as single impressions in the Amgu, Velikaya Kema, and Botchi macrofloras. The coincidence at the generic level is noted only for the Amgu locality, the genus *Osmunda* is identified among impressions and spores. Prevalent are Polypodiaceae spores. As a rule, spores are relatively large, having exine with tubercles and outgrowths unsuitable for transportation over large distances. Spores are preserved better than pollen, as they have a higher content of sporopollenin and lower content of cellulose in their exine (Havinga, 1964; Petros'yants *et al.*, 1990). Kondratene (1976) showed that sporopollenin extracted from *Lycopodium* spores do not suffer from self-oxidation in contrast to sporopollenin of other plants, e.g., of *Pinus* or *Picea*, the oxidized pollen exine of which can be easier destroyed by microbes. This leads to conclusion that presence of spores in the SPS indicates confidently participation of sporogenous plants in vegetative groupings of the region under consideration even in case of their absence in the macroflora collection.

Comparison of generic composition in the micro- and macroflora assemblages shows that the SPS are dominated by pollen of wind-pollinated (anemophilous) plants. This can be explained by their capability to produce a greater amount of pollen and by transportation of the latter over greater distances. A series of works by Zaklinskaya and Grichuk (1948), Malygina (1950), Fedorova (1950, 1952), Fedorova and Vronskii (1980), Kabailene (1969, 1976, 1983), and others demonstrated that greater amount of pollen is transported away from the plant over different distances depending on the plant taxon. Arboreal pollen is transported over larger distances than herbaceous pollen and spores. For

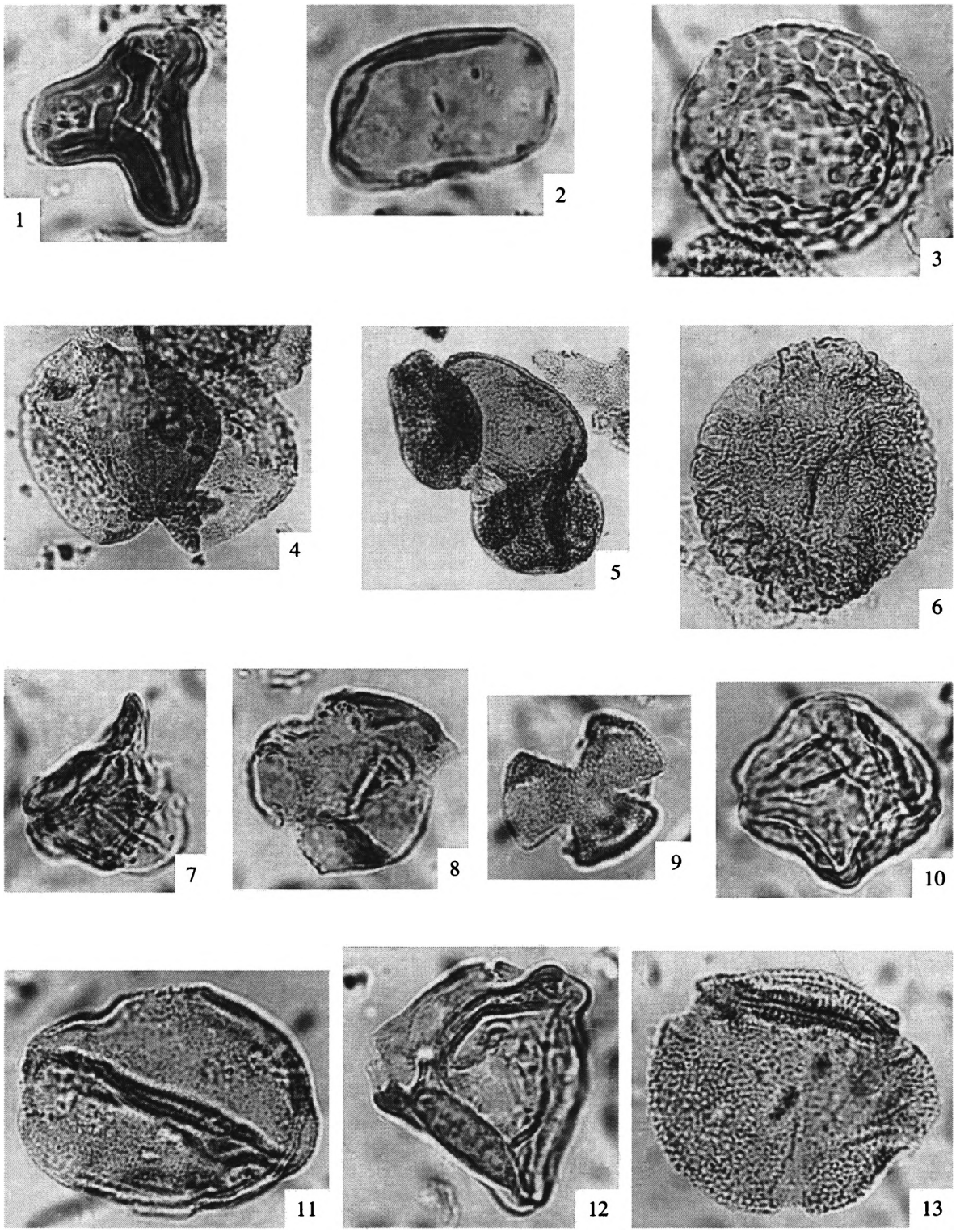


Plate 1. Genera and species of spores and pollen from flora-bearing deposits of the Son'e (upper Eocene), Bui, and Svetlovodnaya (upper Eocene–lower Oligocene) localities: (1) *Cyathea* sp., Sample 3807/18, Son'e flora, ×1000; (2) Polypodiaceae gen. indet., Sample 3807/18, Son'e flora, ×1000; (3) *Botrychium* sp., Sample 3802/1238, Svetlovodnaya flora, ×1000; (4) *Podocarpus* sp., Sample 3807/18, Son'e flora, ×500; (5) *Pinus silvestris* L., Sample 3804/209, Bui flora, ×500; (6) *Tsuga canadiensis* (L.) Carr., Sample 3804/604, Bui flora, ×500; (7) *Cryptomeria* sp., Sample 3807/18, Son'e flora, ×1000; (8) *Cercidiphyllum* sp., Sample 3807/18, Son'e flora, ×1000; (9) *Hamamelis* sp., Sample 3802/1238, Svetlovodnaya flora, ×1000; (10) *Ulmoideipites krempii* And., Sample 3802/1238, Svetlovodnaya flora, ×1000; (11) *Liquidambar styraciflua* L., Sample 3802/1238, Svetlovodnaya flora, ×1000; (12) *Triporopollenites* cf. *plicoides* Zakl., Sample 3802/1238, Svetlovodnaya flora, ×1000; (13) *Fothergilla vera* Lub., Sample 3802/1238, Svetlovodnaya flora, ×1000.

instance, Kabailene (1976, 1983) calculated a radius of the territory, from which a certain quantity of pollen produced by arboreal and frutescent genera have been transported to the observation point under the average wind velocity. It appeared that the transportation radius increases for genera in the following succession: *Carpinus, Corylus, Picea, Fagus, Fraxinus, Quercus, Tilia, Ulmus, Populus, Salix, Alnus, Betula, Pinus*. Kabailene divided genera of studied arboreal and frutescent plants into three groups differing in this parameter. *Pinus, Betula*, and *Alnus* represent a group of the long-range pollen transport. Three forth of pollen produced by these plants has been transported over a distance of 160 to 220 km. *Ulmus, Populus, Tilia*, and *Quercus*, pollen of which (75%) has been transported over a distance of 55 to 100 km, are referred to the group of medium-range pollen transport. The last group of short-range pollen transport (10–22 km) includes *Fraxinus, Fagus, Picea, Corylus*, and *Carpinus*.

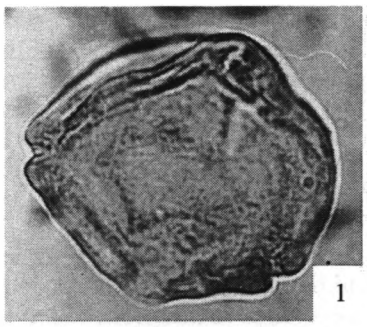
Kabailene (1976, 1983) established as well that palynological spectra of samples collected at different distances from the forest front differ from each other: the farther is the observation point the larger is the source area of pollen. Within the 2-km-wide zone between the forest front and observation point, the radius of pollen source area increases from 21–250 km (the close-range pollen transport) to 280–400 km (the long-range pollen transport), respectively. In the case of sedimentation water basins located in the forest zone, the distance between the observation point and the shore can be considered, with some reservation, as a distance to the forest front (Kabailene, 1983). This seems to be correct for lacustrine sediments considered in this work, and I assume that the studied samples characterize sites adjacent to the forest front, because macroflora remains were evidently buried in the coastal zone of a lake.

Gymnosperms are largely represented, particularly in macrofloras, by Pinaceae (dominant *Picea* and subordinate *Pinus* remains). This is quite understandable in view of the long-range pollen transport and very high pollen productivity of Pinaceae. The long-range transport of *Pinus* pollen is established by Smirnova (1971) who studied recent sediments in the eastern Bol'shezemel'skaya Tundra (Vorkuta region), by Klopotovskaya (1973, 1976) who analyzed subfossil spectra in the Caucasus, and by Borzova (1978) who worked in the eastern Sikhote Alin'. In opinion of Klopotovskaya (1973, 1976), the high content of *Pinus* pollen in spectra not necessary indicates its participation in the phytocoenoses. The forest environments can be suspected when the share of *Pinus* pollen is at lest 90% in the arboreal group representing at least 90% of the total polen spectrum.

Percentages of other Pinaceae (*Abies, Picea, Cedrus*), which produce, like *Pinus*, the bisaccate pollen, are insignificant in the examined preparations, except for samples from the Botchi and Velikaya Kema localities. Concentrations of *Cedrus* and *Abies* pollen in the SPS usually do not exceed 1%, and share of the latter is commonly slightly higher in samples with macroflora impressions. The content of *Picea* pollen is low, except for samples from the Botchi and Velikaya Kema localities where its abundance rate is 31.5 and 20.7%, respectively; percentages of these genera amid macroflora impressions is 6 and 20%, respectively. The *Picea, Abies*, and *Cedrus* pollen is usually larger and heavier than that of *Pinus* and belongs to the short-range transport group. As was mentioned in addition, *Picea* is less productive with respect to pollen than *Pinus* (Kabailene, 1969). Malygina (1950) believes that the content of *Picea* pollen from 10% and higher indicates development of fir forests in the relevant region. In the Eastern Sikhote Alin', the content of *Abies* and *Picea* pollen is 1–2% at a distance of 10–15 km from the source (Borzova, 1978). Proportions of *Picea* and *Cedrus* pollen in spectra can be even lower than the true share of these plants in the vegetative cover of the region. When studying recent alluvial sediments in low mountains of the Lesser Khingan Ridge, Makhova (1971) established that the content of *Picea* pollen in spectra is lower by a factor of 2 to 6 as compared with the true proportion of this species in the forest. *Keteleeria* pollen shows a significant similarity with that of *Abies*, and this complicates identification of this genus especially in preparations with poorly preserved microphytofossils. This property explains probably the absence of *Keteleeria* pollen in macerates from samples with impressions of the genus, which are collected from the Velikaya Kema (Rybalko *et al.*, 1980) and Dembi localities.

It should be noted that all the SPS contain some share of pollen identified as Pinaceae gen. indet., the maximum content of which is established in preparations from the Amgu and Botchi localities. This bisaccate pollen is usually poorly preserved, being deformed, broken, and unidentifiable at the generic level. Musina and Sakhibgareev (1980, 1983, 1984) showed that this pollen experienced physical destruction and fracturing of its exine at the initial stage of sedimentogenesis. The intensity of this process depends on morphology of pollen and spores, granulometric composition of sediments, sedimentation rate, depth and hydrodynamics of the basin. The resistance to destructive physical factors increases from Pinaceae to angiosperm pollen, and then to spores. In particular, pollen of *Abies, Pinus*, and *Picea* is most vulnerable.

Larix pollen is missing from preparations, although the genus is represented by abundant impressions. Two grains presumably belonging to this genus are detected only in a single SPS from the Amgu locality. Such a scarcity of the genus in the polen spectrum reflects a poor preservation of pollen due to its exine structure and low flying capability. The exine of *Larix* pollen is granular, frequently having loose granule arrangement favorable for destruction of pollen grains. The insignificant content of *Larix* in the SPS was noted previously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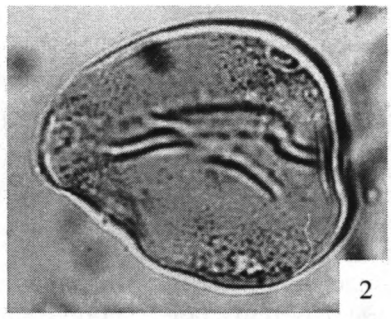
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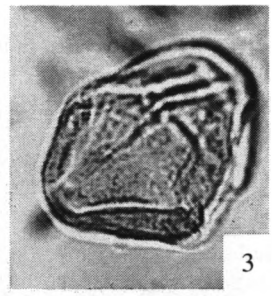
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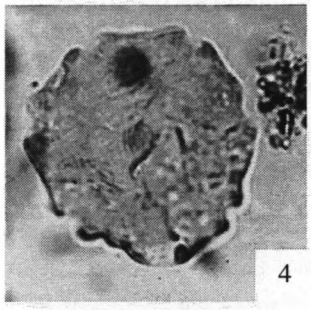
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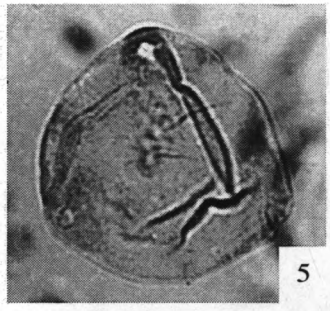
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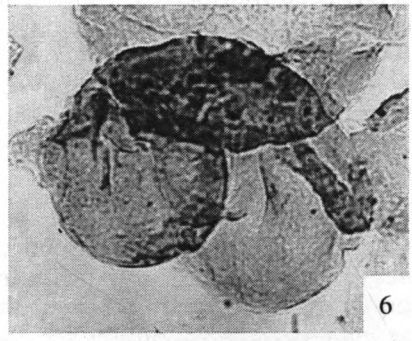
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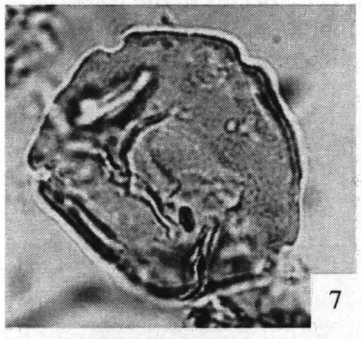
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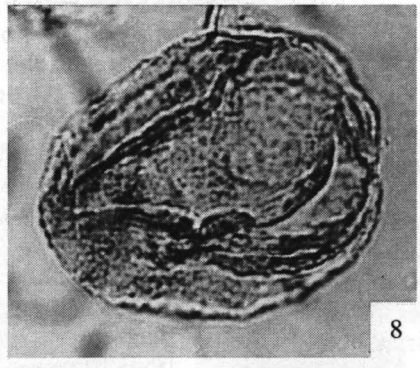
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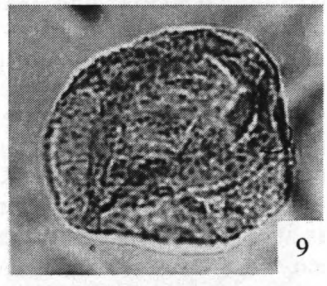
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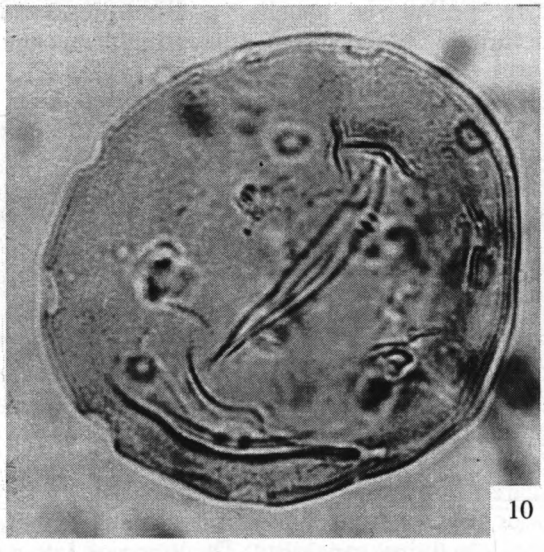
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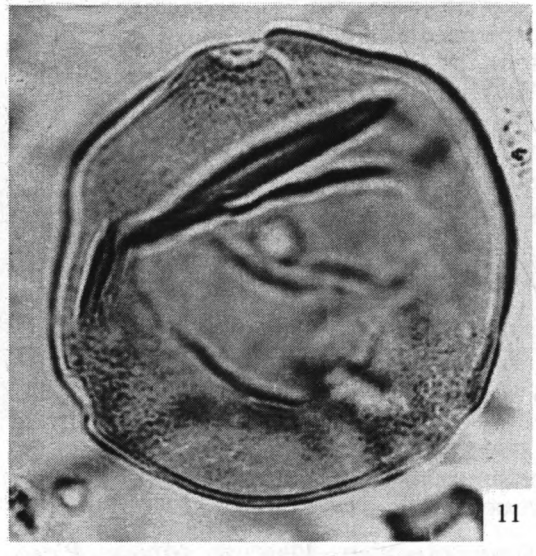
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in many areas with preponderant larch forests where its share is usually below 10%. Kartashova (1971) noted this in recent sediments of the Ola River basin in the northern coast of the Sea of Okhotsk, and Abramova (1978) who studied Pliocene sediments in the southern coast of this sea basin arrived at the same conclusion. The content in spectra up to 1–6% may indicate that more than a half of the relevant area (50–70%) is occupied by larch forests (Makhova, 1971).

Pollen of *Tsuga* is usually well preserved and easily identifiable at the species level, whereas its impressions are commonly suitable for the generic identification. The Botchi SPS includes eight *Tsuga* species, i.e., more than in macroflora collection. Akhmet'ev (1973a) explained this difference as a consequence of similarity between cones of some modern *Tsuga* (e.g. *Tsuga diversifolia*) and larch species (*Larix sibirica, L. leptolepis*), owing to which their discrimination in fossil materials is difficult.

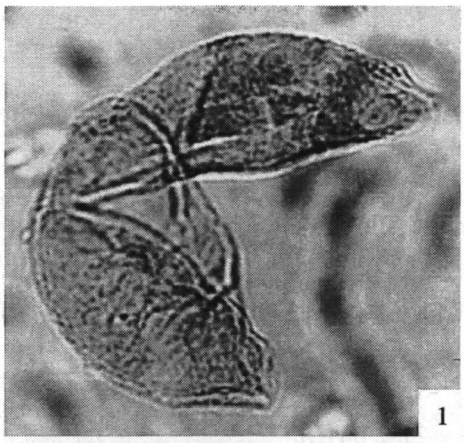
In macroflora collections from the Son'e, Svetlovodnaya, Dembi, and Amgu localities, coniferous are mainly represented by Taxodiaceae and Cupressaceae, percentage of which in the SPS is usually much lower than that of Pinaceae. It is remarkable that *Metasequoia*, which prevails in macroflora collections from the Svetlovodnaya, Amgu, Dembi, and Velikaya Kema localities, is either absent or occurs as rare grains in single SPS. The subordinate role of Taxodiaceae and Cupressaceae amid gymnosperms in fossil palynological assemblages can probably be explained by wrecking of their pollen in distinction from other taxa. Kvavadze (1988) who studied preservation of Taxodiaceae pollen in subfossil and fossil spectra of the Caucasian Black Sea coast demonstrated that pollen of *Cryptomeria, Metasequoia, Cunninghamia*, and *Sciadopitys* is more resistant than that of *Taxodium* and *Sequoia*, because the latter is commonly deformed despite the local conditions favorable for fossilization of spores and pollen (soils with a fine-grained structure and high organic matter content, lacustrine and alluvial sediments). In unfavorable environments, the early-stage fossilization during the first few years results in the partial destruction of the upper exine layer and in rupturing of the pollen aperture area of many genera. Cambon (1994) also noted a poor preservation state of Cupressaceae pollen in recent sediments of Southern Ontario (Canada). Identification of Taxodiaceae and Cupressaceae pollen is often difficult even at the family level. Such unidentifiable pollen usually represents 2–4% of the SPS, but at the Amgu site its content exceeds 10%.

Anemophilous pollen of angiosperms occurs usually in spectra even if these plants are absent in macrofloras. The pollen productivity of angiosperms is commonly high, and their pollen has smooth exine lacking outgrowths that is favorable for its rapid dispersal in the air and for distant transportation. Such pollen is characteristic of small-leaved genera *Alnus* and *Betula*, which are referred, along with *Pinus*, to the group of the long-range pollen transport (Kabailene, 1976), as it was also noted by other researchers (Makhova, 1971; Smirnova, 1971; Klopotovskaya, 1976; Borzova, 1978, and others).

Collections of macroflora and the SPS from the Amgu locality differ in proportions of *Alnus* specimens. This is characteristic as well of the Botchi locality, where *Carpinus* and *Alnus* are dominant. This discrepancy between macro- and microflora assemblages can be explained by the plants habitation in lowlands and low mountains, where they occupy the lower vegetation belt (*Alnus* grows usually along water basin coasts). Therefore, these genera prevail in fossil macrofloras. In the SPS, gymnosperms dominate over angiosperms owing to abundance of Pinaceae pollen, in particular, of *Pinus* pollen. Another factor that can be responsible for the discrepancy is the exine destruction in alkaline soil. Tyuremnov and Berezina (1965) established experimentally that the lower is the pH value (trend to alkaline environments) the easier is destruction of spores and pollen. The so-called thinning of pollen (grains are usually intact, but colorless, having thin exine and indistinguishable structure) is caused by alkalis. One more possible explanation is the capability of *Alnus* and *Quercus* to form hybrids usually having a low pollen productivity.

Pollen of broad-leaved plants (*Carpinus, Juglans, Fagus, Quercus, Ulmus*, and others) is also registered nearly in all SPS. It is of a lower transportation ability in distinction from *Alnus* and *Betula* pollen and can be transported beyond the source area in a negligible quantity. Therefore, its permanent occurrence in spectra implies, even if the content is small, the presence of mentioned genera in phytocoenoses. Klopotovskaya (1985) concluded that spectra of this kind characterize vegetation of small, usually mountain areas where eolian transport of pollen is hampered by orographic barriers. The total share of pollen of broad-leaved *Quercus, Ulmus, Tilia*, and *Corylus* can be however lower by a factor of 2 to 5 than proportion of these threes in a stand, as it was demonstrated for the Amur River basin (Makhova, 1971) and recent sediments of

Plate 2. Genera and species of spores and pollen from flora-bearing deposits of the Svetlovodnaya (upper Eocene–lower Oligocene), Amgu (upper Oligocene–lower Miocene) Dembi, and Velikaya Kema (upper lower–basal middle Miocene) localities: (1) *Carpinus* sp., Sample 3802/1238, Svetlovodnaya flora, ×1000; (2) *Carya* sp., Sample 3802/1238, Svetlovodnaya flora, ×1000; (3) *Rhus* sp., Sample 3802/1238, Svetlovodnaya flora, ×1000; (4) *Pterocarya* sp., Sample 3802/570, Amgu flora, ×1000; (5) *Carya* sp., Sample 3802/570, Amgu flora, ×1000; (6) *Podocarpus* sp., Sample 3803/1194, Dembi flora, ×500; (7) *Juglans sieboldianaformis* Vojc., Sample 3803/884, Dembi flora, ×1000; (8) *Fagus* sp., Sample 3803/884, Dembi flora, ×1000; (9) *Fagus japoniciformis* Anan., Sample 3803/884, Dembi flora, ×1000; (10) *Juglans orbiculata* Bolot., Sample 3820/217, Velikaya Kema flora, ×1200; (11) *Comptonia* sp., Sample 3820/217, Velikaya Kema flora, ×1200.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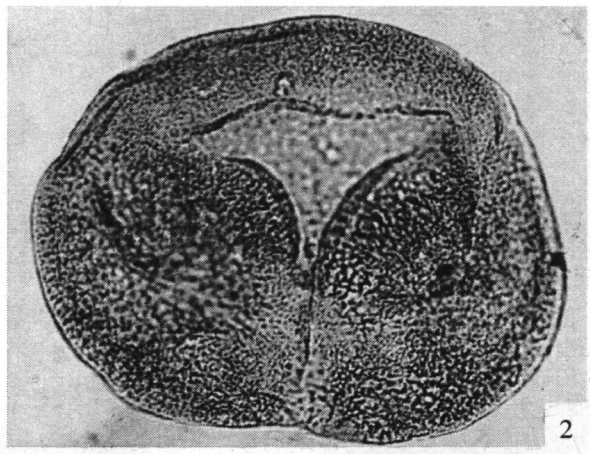
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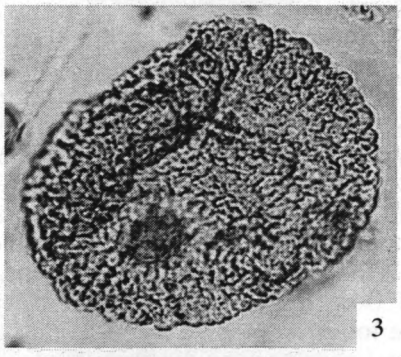
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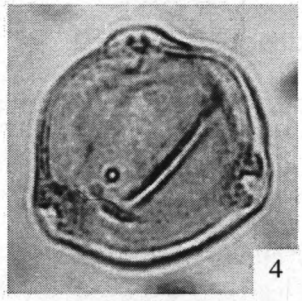
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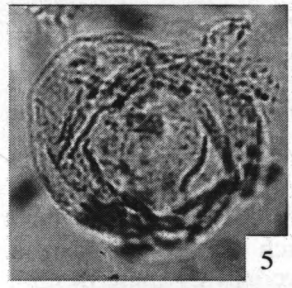
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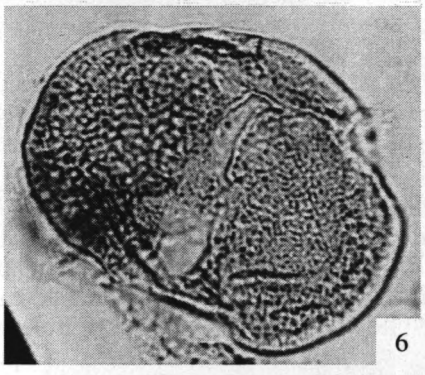
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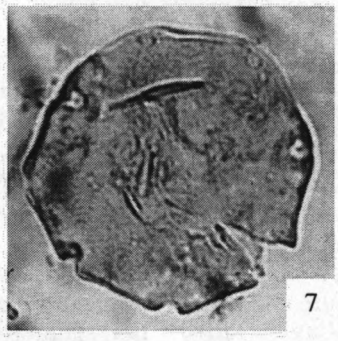
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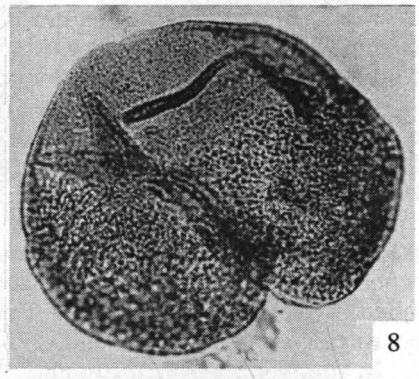
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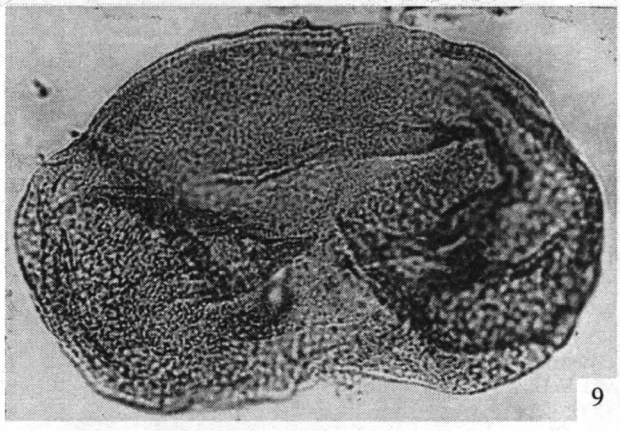
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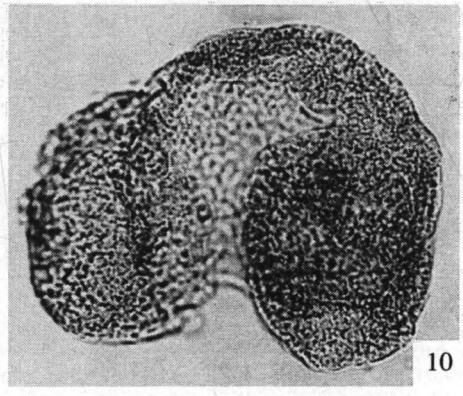
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the peat bogs of Ain, France (Ruffaldi, 1994). In modern beech forests of the Caucasus, the content of beech pollen in spectra from soil and moss samples is several times lower than the plant share in a stand (Klopotovskaya, 1973). In samples from the beech forest, the content of beech and coniferous pollen is 20 and 70%, respectively.

The content of pollen of entomophilous plants (*Salix, Tilia, Castanea, Sterculia, Alangium, Trochodendron, Cornus, Ilex*, and others) is usually low or very low in the SPS; sometimes, it is missing from spectra even if the corresponding impressions are abundant in macrofloras. This holds true for the Bui locality, where the content of entomophilous genera *Tilia* and *Macaranga* is high in macroflora and does not exceed 0.5% in the SPS. This is likely a result of the transport mode of pollen and its morphology. Pollen of entomophilous plants is either large, having an intricate exine ornamentation and, frequently, outgrowths (*Tilia, Sterculia, Diervilla*), or small, united into nubbins (*Salix, Castanea*). As it was shown for contemporary and subfossil assemblages, the content of willow pollen in the SPS is lower than the share of this plant in the vegetative cover (Makhova, 1971; Smirnova, 1971). Even single pollen grains of entomophilous plants indicate their presence in vegetation groupings. Flowers of self-pollinated plants (some species of the *Crataegus* and *Phellodendron*) can remain unblown, and the plants release their pollen, if any, in an insignificant quantity (Fedorova and Vronskii, 1980).

Pollen of most herbaceous plants is either scarce in or missing from the SPS. When present, it belongs usually to aqueous plants. Several factors can be responsible for this situation. First, the transport distance of herbaceous pollen is shorter than that of arboreal taxa. Second, a characteristic feature of many herbaceous plants is their ability of cloning. Third, the herbaceous layer can be poorly developed because of the sun light deficiency in shady forests (Zaklinskaya, 1950; Fedorova and Vronskii, 1980). In the case of underwater pollination, proper plants produce pollen with a very thin underdeveloped exine that can be easily destructed. Moreover, the pollen productivity of aqueous plants is usually lower than that of land vegetation. The low content of impressions of herbaceous and aqueous plants in macroflora collections can be explained by their insufficient resistance to unfavorable burial conditions and *in situ* destruction.

It should be noted that pollen of *Acer, Populus*, and Rosaceae is poorly resistant because of a low sporopollenin content in exine easily destructible therefore during oxidation (Vronskii and Fedorova, 1981; Sangster and Dale, 1964). Exine of *Populus* and Rosaceae pollen is very thin and fragile, and their specimens are extremely scarce in the SPS. *Populus* is missing from all examined preparations, although impressions of this plant are present in some samples. Impressions of Rosaceae are relatively diverse in macroflora collections from the Dembi and Amgu localities, but their pollen is absent in the SPS. Some of tricolporate pollen grains of uncertain taxonomic affinity can probably represent this family. The content of *Acer* pollen never exceeds 0.5%, whereas impressions of this genus are abundant in the Bui, Svetlovodnaya, and Dembi macrofloras. The Amgu SPS contains about 4% of poorly preserved pollen belonging to this genus.

Some Betulaceae, Juglandaceae, and Hamamelidaceae pollen grains are unidentifiable at the generic level. The share of these families in spectra is usually below 1%, and their pollen could be decomposed at the stage of sedimentation.

In some samples entirely barren of spores and pollen, palynomorphs could be demolished during diagenesis. At this lithogenetic stage, destruction is of a chemical–biological type, and Petros'yants (1990) reported that precisely this process is responsible for occurrence of continental and marine sediments lacking palynological remains. The diagenetic destruction results in appearance of cavities, dotty harms, and branching damages, the first two typical of the ectoexine and the third ones characteristic of the endoexine. The damaging degree depends on the exine structure and chemical composition, and also on the sediment permeability and on the fluid pH. I distinguished dotty damages on spore and pollen grains and on algae remains (*Botryococcus* and *Ovoidites* sp.) in samples from the Svetlovodnaya locality. Cavities are characteristic of palynomorphs from the Son'e, Amgu, and Dembi (four samples) localities, and exine thinning is registered in samples from the Botchi (three samples) and Dembi (two samples) localities. These damages point to the chemical–biological destruction of spores and pollen during their fossilization. Most strong damages have been noted on pollen grains of coniferous and angiosperm plants.

As is mentioned, the examined plant remains were buried in lacustrine sediments. Many works dedicated to the problem of burial of palynological remains in lakes (Matveeva, 1950; Kabailene, 1969; Malyasova, 1980, 1984; Kleimenova and Khomutova, 1979; and others) showed that pollen is distributed more or less

Plate 3. Genera and species of spores and pollen from flora-bearing deposits of the Velikaya Kema (upper lower–basal middle Miocene) and Botchi (uppermost middle–upper Miocene localities: (1) *Thuja* sp., Sample 3820/217, Velikaya Kema flora, ×1200; (2) *Picea excelsa* L., Sample 3820/217, Velikaya Kema flora, ×500; (3) *Tsuga macroserrata* (Wolff) Anan., Sample 3805/5, Botchi flora, ×500; (4) *Betula* sp., Sample 3820/217, Velikaya Kema flora, ×1200; (5) Taxodiaceae or Cupressaceae, Sample 3805/2, Botchi flora, ×1000; (6) *Pinus tertiaria* (Moreva) Anan., Sample 3805/1, Botchi flora, ×750; (7) *Juglans* sp., Sample 3805/1, Botchi flora, ×1000; (8) *Picea media* Anan., Sample 3805/5, Botchi flora, ×500; (9) *Picea* cf. *obovata* Led., Sample 3805/5, Botchi flora, ×500; (10) *Pinus* sp., Sample 3805/1, Botchi flora, ×500.

regularly over the flat bottom of shallow lakes. In deep lakes with differentiated bottom topography, the SPS compositions are dependent on bathymetric positioning in the coastal shoal, transitional or slope and deep zones of a basin.

Kabailene (1969) demonstrated that the highest content of palynological remains is characteristic of the deep zone. The content of *Pinus* pollen decreases from the coastal shoal toward the deep zone, whereas the abundance growth in the same direction is characteristic of *Picea, Betula, Alder, Corylus, Carpinus, Quercus, Tilia, Ulmus* and herbaceous pollen and spores. Davis and Brubaker (1973) also noted the high content of *Pinus* pollen in shallow-water coastal sediments.

Pollen of *Pinus* usually prevails among gymnosperms in the all examined SPS, and this likely a taphonomic peculiarity and consequence of factors mentioned above (high pollen productivity, pollen morphology, and long-range transport). Leafs have been buried in a shallow coastal zone of the lake, where the high content of *Pinus* pollen is recorded.

CONCLUSION

Examination of macro- and microfloras from seven localities of the Eastern Sikhote Alin' shows that they bear a noticeable similarity at the generic and family levels. The following genera usually present in both floras: *Pinus, Picea, Taxodium, Glyptostrobus, Sciadopitys, Thuja, Carpinus, Ostrya, Juglans, Carya, Fagus, Quercus, Ulmus, Zelkova, Cercidiphyllum, Rhus,* and *Ilex.* The palynological method reveals a greater number of genera. For instance, genera of sporogenous and aqueous plants, some representatives of Hamamelidaceae and Araliaceae, and genera *Magnolia, Myrtis, Loranthus*, and have been identified only in palynological spectra. In contrast, Rosaceae and genera *Populus, Diospyros, Styrax, Ligustrum*, and some others are found only among plant impressions. The qualitative and quantitative differences between particular assemblages reflect impacts of several factors: (1) biological peculiarity of plants (chemical composition of exine that controls preservation of spores and pollen, pollination mode, variable productivity of spores and pollen, and their morphology, flying capability, buoyancy, and specific weight); (2) peculiarities of micro- and macroflora fossilization; (3) regional physiographic conditions (topography of adjacent areas and lake parameters such as depth, bottom topography, coastal and deep currents); (4) lithogenetic conditions (type and alteration degree of enclosing sediment); (5) allochthonous and autochthonous genesis of microfossils and macroflora, respectively; (6) insufficient knowledge of morphology of some spores and pollen taxa).

ACKNOWLEDGMENTS

I express my sincere gratitude to M.A. Akhmet'ev (Geological Institute RAS), I.A. Basov, and A.Yu. Gladenkov (Institute of the Lithosphere of Marginal Seas RAS) for discussions during preparation of the manuscript, to A.I. Nazarov (Geological Institute RAS) for his help in photographing, and to N.G. Tremasova (Institute of the Lithosphere of Marginal Seas RAS) for drawing figures. The work was supported by the Russian Foundation for Basic Research (projects no. 99-05-65604 and 01-05-06100).

Reviewers M.A. Akhmet'ev and N.I. Zaporozhets

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Stratigraphy and Geological Correlation, Vol. 11, No. 1, 2003, pp. 91–103. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 11, No. 1, 2003, pp. 96–110.
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The Pleistocene Glaciation Rhythms

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Received November 27, 2000, in final form, December 3, 2001

Abstract—Our own and published paleogeographic data have been revised to reliably define rhythms and intensity of climatic events in the glaciation areas of the Russian Plain during the Brunhes Chron. Biostratigraphic and lithostratigraphic studies enable recognition of till horizons and other diagnostic features of seven Neopleistocene glaciations. The extreme distribution limits of morains are specified. The distribution areas of corresponding glacial sheets gradually decreased in the Russian Plain since the Don to the late Valdayan Glaciation time. The independent status of the Dnieperian and Moskowian glaciations remains a matter of controversy. Some stages of vegetation development in the central Russian Plain are outlined. Paleobotanical data have been used to reveal the short–term and longer climatic rhythms. These data also show that the Akulovo Interglacial is an Early Eopleistocene event. Complex paleogeographic patterns of the post-Don interglacial epochs are described. The established comparative intensity of climatic events in the corresponding latitudinal zones of continent and oceans has been used to correlate the Valdayan, Moskowian, Dnieperian, Okaian, Don, and Setun' glacials with the oxygen-isotope stages 2–4, 6, 8, 12, 14, and 16, respectively. Correlation between shorter-term climatic events in continental and oceanic regions is detectable in the post-Don period.

Key words: Neopleistocene, Russian Plain, paleogeographic reconstructions, vegetation stages, glaciation rhythm, correlation between continental and oceanic climatic events.

INTRODUCTION

After the work by A. Penck and E. Bröckner published in 1909, four great glaciations separated by interglacial epochs have been considered as characteristic of the Pleistocene. Analogous of these glaciations (including the fifth Donau glaciation that was established in piedmonts of the Alps 20 years later) were identified in plains of Western Europe, Russia, and North America. The glaciation periods were subdivided into stages and substages.

Development of oxygen-isotope chemostratigraphy after the early study of Emiliani and Shackleton (1974) resulted in recognition of 19 cooling and warming events during the Brunhes Chron, which are recorded in oceanic sediments. Based on the high-resolution oxygen-isotope and foraminiferal analyses of continuous oceanic sequences, Van Donk (1976) found out that the isotopic effect of temperature fluctuations did not exceed 20% in the most cases, and that the oxygen isotope variations in tests of planktic foraminifers reflected predominantly the secular changes in ice volume on the Earth.

These data correspondingly changed viewpoints on continental glaciations. Researchers tried to discover diagnostic features of nine continental glaciations of the Brunhes Chron in accordance with the quantity of even isotope stages indicative of cooling. However, up to recently there are no convincing records of nine continental till horizons of different age. The number of glaciations in so well studied region as the Russian Plain is far from being well established, as it is evident by publications of the last decade (Velichko *et al.*, 1989; Sudakova, 1990; Zarrina, 1991; Velichko and Shik, 1992; Zubakov, 1992; Shik, 1993; Alekseev *et al.*, 1997; Andreicheva *et al.*, 1997; Krasnenkov *et al.*, 1997; and others). Moreover, the glacial horizons shown in stratigraphic schemes have not been all substantiated by factual data. Some of them are products of theoretical considerations of the authors.

Still problematic are climatic interpretations of some interglacials, ranking of individual warming events, and their stratigraphic intervals (Velichko *et al.*, 1989; Grichuk, 1989; Zubakov, 1992; Pisareva, 1997; Krasnenkov *et al.*, 1997; Velichko *et al.*, 1999; and others). In general, the progress in oceanic oxygen-isotope investigations repudiated importance of glacial events ranking probably because of relatively uniform amplitudes of corresponding isotopic signals.

Available inadequate methods of age determination of terrestrial and oceanic Quaternary deposits led to development of "astronomic stratigraphy" (Johnson, 1982; Shackleton *et al.*, 1990; and others). The harmonic analysis showed that cyclic changes in the Earth orbital parameters (with periods of about 100, 41 and 23 kyr) could control amplitudes of Pleistocene climatic variations nearly to the level of 60%. The corre-

lation between cyclic changes of $\delta^{18}O$ values in calcareous foraminiferal test and the Milankovitch cycles triggered the age recalibration of oxygen isotope stages. According to the currently accepted dates (Bassinot *et al.*, 1994), the Brunhes–Matuyama boundary is defined at the level of 780 ka. However, opinions about that boundary position in individual sections are controversial. Different authors placed it inside the Mikhailovian (Petropavlovsk) Horizon (Bylinskii, 1992), at the levels of Mikhalovka–Platovka (Nikiforova, 1982) or Petropavlovsk–Pokrovskii (Velichko and Shik, 1992) boundaries, in the upper part of the Petropavlovsk Horizon (Shik, 1993; Vangengeim *et al.*, 2001), and even inside the Il'inskii Horizon.

A greater discrepancy is between the northern glacial and southern loess–soil sequences. In the Upper Don basin, the Don till and up to three soil horizons occur between the Matuyama–Brunhes boundary (the supposed position) and the Vorona Soil, whereas in the southern sections the Matuyama–Brunhes boundary is defined in the Don loess and, therefore, a noticeable stratigraphic interval is missing. Thus, the stratigraphic position of the Matuyama–Brunhes boundary is uncertain, placed at different regional levels (Faustov *et al.*, 1997).

Its position is, however, of crucial importance for correlation of continental and oceanic climatic events. The commonly used method of correlation by "counting events from the top down" frequently yielded different results because of different interpretations of continental sequences.

For example, glacial rhythms can be correlated with odd isotopic stages, as it has been done by Velichko *et al.* (1989) who attributed the Don glacial to oxygen isotope stage 19. In other cases, several climatic rhythms may be correlated with a single isotopic stage as in the scheme by Shik (1993) who correlated two Lower Neopleistocene subdivisions of a complex structure (the Il'inskii and Pokrovskii horizons) with the stage 18. Moreover, the Il'inskii Horizon of that scheme includes two pre-Don glacial horizons, although Zubakov (1992) and Zarrina (1991) attributed the first of them to the Pokrovskii Horizon. Examples of this kind are numerous.

Based on published data and results of our long-lasted studies in oceans and on continents, we intend to revise the available materials in order to select the most reliable paleogeographic reconstructions. These reconstructions and newly calibrated oxygen isotope scale will be used to correlate oceanic and continental paleoclimatic events of the first order and to evaluate the possibility of recognizing the more frequent oscillations.

MATERIALS AND METHODS OF INVESTIGATION

The considered materials include our results of long-lasted field observations with subsequent laboratory work in the glaciation area of the Russian Plain (Fig. 1) and published data on terrestrial and oceanic regions. We summarized representative results of lithological, mineralogical, petrographic, carpological, palynological, and diatom analyses and supplementary paleontological, geochronological, and paleomagnetic data on continental deposits. A special attention was paid to statistic processing of oxygen isotope variations in foraminiferal tests and oceanic sediments.

Since our own data mostly characterize continental sequences of the East European plain, we concentrated our attention on development of the Scandinavian glacial sheet that is supposed to have been concurrent with development of glaciation in North America during the Pleistocene time.

In this work, we share the classical views of Markov (1955) on glacial and interglacial epochs. We considered data on the past glaciation areas leaving aside many supplementary materials on concurrent non-glacial deposits. Such an approach, limiting possibilities of the analysis on the one hand, is most convenient for paleoclimatic interpretations of true till horizons and strata separating them, on the other.

The work included three interconnected procedures: (1) summarization of climatic, lithological, and stratigraphic records in the past glaciation areas, (2) paleolandscape analysis of central zone in the East European plain, and (3) correlation between terrestrial and paleoceanological records.

RESULTS AND DISCUSSION

Climatostratigraphy of the glaciation area. A complex cyclicity of the Early Neopleistocene events is established in the central Russian Plain. At least three interglacial stages are distinguished here with confidence: the Akulovo and Okatovo interglacials predating the Don glacial and the Roslavl' (Muchkap) interglacial after the latter (Maudina *et al.*, 1985; Fursikova *et al.*, 1992; Shik, 1993; Pisareva, 1997).

The earliest paleobotanical studies of the oldest Akulovo interglacial deposits were carried out near the Akulovo Village, the Odintsovo Station of the western Moscow region (see section localities in Fig. 1). Abundant and diverse fossil flora obviously inherited from the Pliocene implies that the deposits should be attributed partly to the Eopleistocene and partly to the basal Neopleistocene.[1] Distinctive features of the Akulovo flora are as follows: a high diversity of conifer and broad-leaved forms, including the most thermophilic *Eucommia, Castanea, Pterocarya, Morus, Ilex*, and other genera; abundance (more than 30%) of local and regional exotic taxa; and frequent (more than 11%) occurrence of extinct species *Aracites* cf. *interglacialis*

[1] Stratigraphic range of the Neopleistocene is accepted in accordance with the resolution of Interdepartmental Stratigraphic Committee, February 2, 1995, and approximately corresponds to the Brunhes Chron. The Eopleistocene is equal to the Apsheronian Stage, which was previously referred to the upper Pliocene.

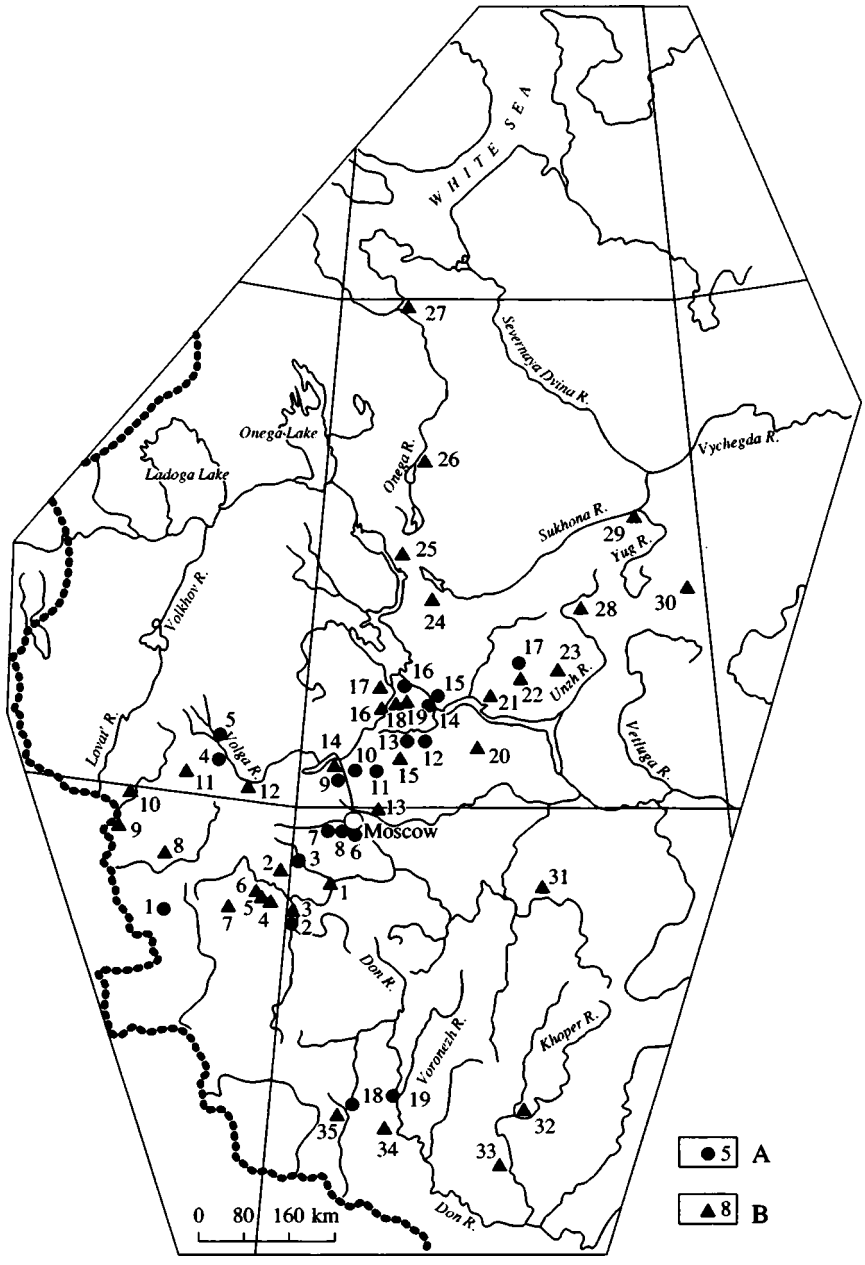

Fig. 1. Main sections of glacial and interglacial sediments in the Russian Plain; (A) key areas and sections by names: (1) Roslavl' (Konakhovka, Malakhovka, Sergeevka), (2) Chekalin (Likhvinian), (3) Borovsk–Satino, (4) Molodoi Tud, (5) Manukhino, Loshakovo, (6) Moscow, (7) Odintsovo–Akulovo–Okatovo, (8) Lytkarino, (9) Spas-Kamenskii, (10) Dmitrov, (11) Kun'ya, (12) Rostov–Nero, (13) Cheremoshnik–Shchurskol, (14) Yaroslavl–Pechegda, (15) Tutaev–Dolgopolka, (16) Rybinsk–Chermenino–Yakovka, (17) Galich–Gorki, (18) Stoilinskii, (19) Strelitsy; (B) other sections: (1) Tarusa, (2) Myatlevo, (3) Bryan'kovo, (4) Meshchevsk, (5) Mosal'sk, (6) Resseta, (7) Spas-Deminsk, (8) Yartsevo, (9) Kasplya, (10) Smolenskii Brod, (11) Koslovo–Bereza, (12) Vazuza–Zubtsov, (13) Balashikha, (14) Mar'ino–Pan'kovo, (15) Lake Pleshcheevo, (16) Altynovo, (17) Shestikhino, (18) Uglich, (19) Bol'shoe Selo, (20) Bibirevo, (21) Zakhar'ino, (22) Chelsma, (23) Pepelovo, (24) Molochnoe, (25) Ferapontovo, (26) Kargopol', (27) Onega, (28) Anyug, (29) Velikii Ustyug, (30) Gorokhovskaya, (31) Narovatovo, (32) Novokhopersk, (33) Nizhne-Dolgovskii, (34) Uryv, (35) Lebedinskii.

Nikit., *A.* cf. *jonstrupii* (Hartz) Nikit., *Azolla interglacialis* Nikit., and some others (Maudina *et al.*, 1985; Pisareva, 1997).

Stratigraphic range of the Akulovo Interglacial was considered earlier (Pisareva, 1997) in view of discrepancy between paleomagnetic and palynological records. Based on paleomagnetic data, V.V. Semenov attributed the Akulovo deposits to the Brunhes Chron (see in Maudina *et al.*, 1985), whereas our paleobotanical data indicate its older age.

The Brunhes–Maruyama boundary has not been reliably defined either in the central part of the glacia-

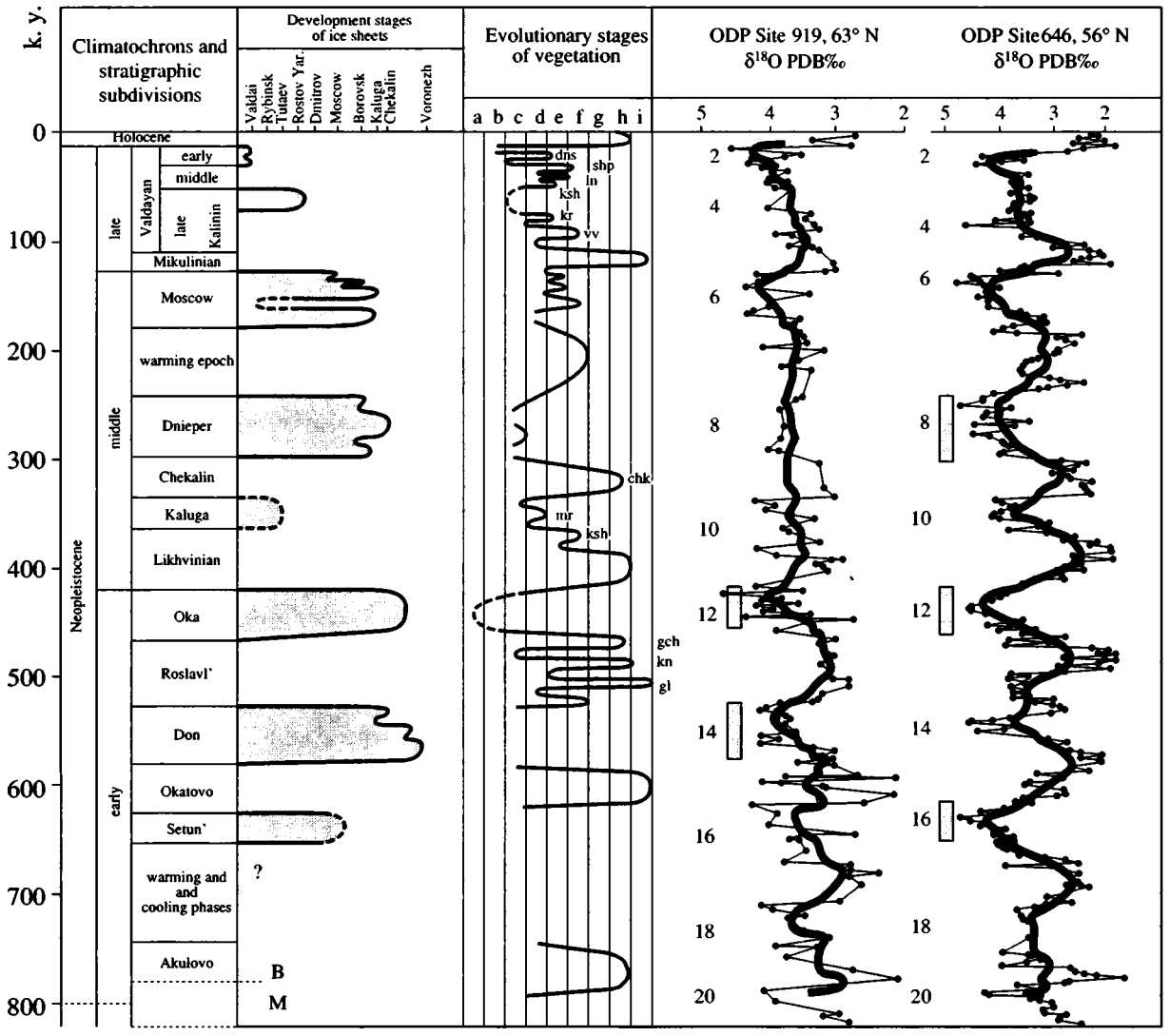

Fig. 2. The Pleistocene paleogeographic rhythms; the reliably established and presumable glaciation and deglaciation epochs are contoured by solid and dotted lines respectively; stages of vegetation evolution in the Russian Plain are outlined for a zone between 56° and 57° N. Landscapes: (a) arctic desert, (b) tundra–steppe, (c) periglacial forest steppe, (d) light conifer and small-leaved forests, (e) northern taiga, (f) intermediate taiga, (g) southern taiga, (h) coniferous–broad-leaved forests, (i) broad-leaved forests. Climatostratigraphic units of subordinate rank: (gl) Glazovo, (kn) Konakhovo), and (gch) Galich optima; (kch) Koshino, (mr) Mar'ino, and (vv) Verkhnyaya Volga interstadials; (chk) Chekalin, (kr) Krasnogorsk, (ksh) Kashino, (ln) Leningrad, (shp) Shapurovo, and (dns) Dunaevka warming phases. Curves depicting the high-frequency and major secular variations of $\delta^{18}O$ in calcareous tests of planktonic foraminifers from the ODP sites in the Atlantic Ocean are shown by thin and thick lines respectively (after Nikolaev, 2000). Shaded rectangle indicates the most significant climatic events of the early–middle Pleistocene, which are discussed in the text. The thick dotted line denotes the supposed position of the Brunhes (B)–Matuyama (M) boundary based on paleobotanical data.

tion area of the East European Plain or in the Baltic region (Kondratene, 1996). At present it is established at the level of 780 ka (Shackleton *et al.*, 1990; Bassinot *et al.*, 1994) near the Neopleistocene–Eopleistocene boundary. The greater part of the Akulovo interglacial (Fig. 2) and preceding cooling event (the problematic Likova glacial) likely correspond to the Eopleistocene. According to Zagwijn (1996), the diverse flora with *Eucommia* resembling the Akulovo flora is recorded only in the lower Cromerian horizons (Interglacial 1) corresponding to the Matuyama Chron. In the Baltic region, the taxonomically similar flora was found in the Eopleistocene deposits of the Daumantai section of southeastern Lithuania (Kondratene, 1996).

In distinction from the Akulovo deposits, the Okatovo interglacial deposits (reference sections of the synonymous site in the western Moscow region) contain less diverse conifer and broad-leaved forms. In opinion of T.V. Yakubovskaya who studied plant impressions (see in Fursikova *et al.*, 1992), the Okatovo flora is most similar to the early Pleistocene flora of the Korcheva

section in Belarus.[2] Erratic boulders of the Okatovo gyttja frequent in the Don till suggest a broad distribution of corresponding lacustrine–boggy deposits in the pre-Don time (Shik and Pisareva, 1998). Fossil floras indicate that the Akulovo and Okatovo interglacials were separated by a long interval of alternating cooling and warming events of different ranks.

As for the earliest pre-Don glaciation areas, they are unclear so far. Relationships between the pertinent till horizons and their stratigraphic positions have not been well substantiated, and lithological composition of boulder deposits is adequately studied. The earliest glaciations are inferable only from fragmentary indirect evidences (erratic pebblestones and so on).

The oldest Likova till distinguished at the base of the Akulovo section in the Moscow region (Maudina *et al.*, 1985; Shik, 1993) seems to be suspicious, because that thin bed of boulder loam is of problematic genesis. The first Setun' till is the only one confidently predating the Don glaciation in the Moscow and Roslavl' areas, though it is established in isolated sections based on the relative stratigraphic position and has not been traced elsewhere. Other sporadic occurrences of erratic pebblestones suggest that the pre-Don glaciations did not advance southward of northern and central areas of the Russian Plain.

For this reason, the extension of the Setun' glacier cover to the south of Moscow is shown quite arbitrary in Fig. 2 illustrating the inferable paleogeographic rhythms of the Neopleistocene. However, cooling at the Setun' time was significant enough to result in permafrost, as it is evident from wedge-shaped polygonal structures in corresponding sediments of the Don and Dnieper river basins (Velichko *et al.*, 1999).

The marker horizon of three-layer Don till is traceable in the Oka–Don and Central Russian plains, and also on slopes of the Central Russian and Volga uplands. The horizon time span is reliably restricted by occurrence of rodent bones of the middle Tiraspol' age in the underlying Moiseevo Formation (the Muchkap locality) and by the late Tiraspol' fauna of small mammals from the overlying alluvial deposits (Krasnenkov *et al.*, 1997). The late Tiraspol' fauna was found above the Don Till in many other sections of the Don (Vol'naya Vershina, Korostelevo, Kuznetsovka, Klepki, and others) and Dnieper (Konakhovka near Roslavl', Podrudnyanskii, Segeevka, and others) river basins (Krasnenkov *et al.*, 1984; Agadzhanyan, 1992; Markova, 1998). It indicates the early Neopleistocene age of the Don Till. In the western Moscow and Smolensk areas, the Don Glaciohorizon underlying the Roslavl' interglacial deposits is known under names of Pereksha or Oster tills (Maudina, 1985; Biryukov *et al.*, 1994).

In the Don River basin, the synonymous glaciohorizon is of specific composition (Gribchenko, 1980; Krasnenkov *et al.*, 1984; Runkov *et al.*, 1993) and exhibits various glaciotectonic structures, forming outliers especially widespread in the southwestern part of its distribution area. The mineral composition of the till matrix, which was studied recently in the Strelitsa, Stoilinskii, Novokhopersk, Narovatovo and other key sections, and the petrographic composition of boulders are obviously interconnected with the underlying deposits. Main heavy minerals, such as disthene, glauconite, sulfides, and tourmaline, have been derived from local provenances and associate with a minor (4 to 6%) admixture of exotic hornblende.

Although the Don till has been comprehensively studied, separate stages of the Don glaciation, which are recorded in its three-layer structure with essential mineralogical and petrographic differences between the layers, are inadequately understood. The genetic relation of these persistent till beds to different glaciation stages is also evident from the fluvioglacial origin of deposits separating them. The glaciation center cannot be defined with confidence in the stratotype area, because it is impossible to make a correct choice between the Scandinavian or North Uralian provenances of erratic material.

The succeeding Roslavl' (Muchkap) Interglacial is among the best studied ones. The deposits overlying the Don till contain rodent fauna (Velichko *et al.*, 1989; Agadzhanyan, 1992; Markova, 1998) and abundant floral remains (Velichkevich, 1982; Krasnenkov *et al.*, 1984; Biryukov *et al.*, 1994; Pisareva, 1997). During the Roslavl' Interglacial there were three climatic optima (Maudina *et al.*, 1985; Pisareva, 1997). The Glazovo and Konakhovo optima are recorded in palynological diagrams from the Roslavl' stratotype region and Moscow area (Shik *et al.*, 1998). The third Galich optimum was established for the first time in the glacial sequence near the town of Balashikha (Pisareva *et al.*, 1979) and later on in the sections near the Akulovo Village and Odintsovo Station, and in the Galich area at the Chelsma River. Intervals between the optima characterize impoverishing of floral assemblages, but they do not yield floral or faunal evidences indicating the ice sheet development in proper times. Cryogenic structures have not been also discovered in these intervals.

The early Neopleistocene age of the Roslavl' interglacial deposits is also proved by their position under the Oka till and Likhvinian interglacial deposits in the Roslavl' stratotype region (the Malakhovka locality) and other sections, e.g., westward or northward of Moscow and in the Dnieper River basin (Shik and Pisareva, 1998).

In general, the Roslavl' deposits are well correlative with those of the Byelovezhian Interglacial of Belarus (Voznyachuk, 1985), the Ferdynandow Interglacial of Poland (Janczyk-Kopikowa, 1991), and the Voigtstedt Interglacial of Germany (Erd, 1972). The same event is recorded in upper horizons of the Cromerian Intergla-

[2] Voznyachuk (1985) believed that the Korcheva interglacial predates the Günz epoch and is much older than the Byelovezhian interglacial.

cial of East England (West, 1980) and Holland (Zagwijn, 1996).

The next Okaian glaciation of the early Neopleistocene age can be correlated with the Elsterian glaciation of Western Europe. Its traces are confidently established in the Likhvinka, Bryan'kovo (*Razrezy...*, 1977) and Tarusa (Sudakova, 1990) sections of the Oka River basin, in the Roslavl' stratotype region (Biryukov *et al.*, 1994), and in the Narovatovo key section (Runkov *et al.*, 1993). Everywhere the Oka till underlies the Likhvinian interglacial deposits.

An interval of reversed magnetic polarity established in the Bryan'kovo section and known as synonymous paleomagnetic excursion (Sudakova and Bol'shakov, 1977; Isaeva *et al.*, 1978) allows the Okaian glacial sediments to be correlated with the Dainava till (Elsterian) of Western and Central Europe and of Lithuania (Gaigalas, 1979). Although the Oka till is exposed along the right bank of Oka River in its middle courses down to the lower course of the Moksha River, its distribution limits need to be specified. The Oka till, in which erratic material of local origin dominate over that from remote provenances, is similar to the Don till in petrographic and mineralogical characteristics. However, a distinctive feature of the former is higher percentages of Fennoscandian hornblende (~20%), garnet (~15%), epidote (~15%), zircon, and other accessory minerals of the heavy fraction.

The Likhvinian Horizon of the middle Neopleistocene interglacial deposits is well correlative to the Holsteinian interglacial deposits of Europe based on both the floral and faunal remains (Grichuk, 1989; *Razrezy...*, 1977; Shik and Pisareva, 1998; Markova, 1998; and others). After a thorough study, vegetation and climate stages of the interglacial have been specified (Fig. 2).

By now the succession of warm and cold phases of the entire Okaian–Dnieperian interval has been interpreted a new. In sections of the Tver' (at the Bol'shaya Kosha River, near the Bulatovo and Tyaglitsy villages) and Moscow (at the Pan'kovo and Mar'ino villages) regions, there are established two, the Kosha and Mar'ino interstadials postdating the Likhvinian Interglacial (Pisareva, 1997). The latter was succeeded by the significant Kaluga cooling and Chekalin warming events, the latter established for the first time in the Likhvinian (Chekalin) stratotype section (Sudakova, 1975; Bolikhovskaya, 1995).

The Kosha and Mar'ino interstadials may be analogous to the Hoogeveen and Bantega interstadials established by Zagwijn (1985) above the Holsteinian Interglacial in the Dutch sections. One of these interstadials is shown in the middle Neopleistocene stratigraphic scheme of France (Reille *et al.*, 1998). The mentioned authors also recognized the next Bargette Glacial, probably corresponding to the Kaluga cooling event, and the well-pronounced Landos Interglacial likely correlative with the Chekalin warming phase in the Russian Plain. In German sections, the Fuhne Glacial and Dömnitz Interglacial, or "Waken warm time," are recorded above the Holsteinian Interglacial (*Das Quartär Deutschland*, 1995).

The middle Neopleistocene glacial sequence includes two, Dnieper and Moscow tills deposited during the great and independent Dnieperian and Moskowian glaciations, as Sudakova believes.[3] Type sections of both glaciohorizons are known in the Moscow region (middle courses of the Protva River) and in the Rzhev and Yaroslavl' areas of the Volga River basin. These horizons have been correlated with the Pechora and Vychegda tills of the northeastern Russian Plain based on lithological criteria (Andreicheva *et al.*, 1997).

The stratigraphic position of the Dnieper till is clearly defined in the Chekalin and Rybinsk key sections (*Razrezy...*, 1977; Sudakova, 1990; Agadzhanyan, 1992; Sudakova *et al.*, 1996). Its diagnostic features are as follows: predominant orientation of elongated clasts toward the Timan–Urals region, the former provenance of erratic material (Karpukhin and Lavrov, 1974; Sudakova *et al.*, 1995; Andreicheva *et al.*, 1997), and specific petrographic and mineral compositions.

In distinction from the Dnieperian Glaciohorizon, the Moscow till is depleted in epidote and garnet and essentially enriched in hornblende, the leading mineral (25 to 30% of heavy fraction) of the Fennoscandian provenance. This change in mineral composition accords well with facts suggesting the southeastward glacier movement from the Scandinavian glaciation center.

The Moscow glaciation was responsible for appearance of characteristic relief forms all along its front from the Luzha, Protva and Volga rivers up to the Yaroslavl area (*Moskovskii...*, 1982; Sudakova, 1990). The interstadial deposits were discovered in the Kostroma (Pisareva, 1965) and northern Moscow (Nosov and Skiba, 1975) areas, and lately in upper courses of the Protva River. However, correlation of these deposits based on paleobotanical data can be suspicious, especially when they are characterized fragmentarily or their stratigraphic position is insufficiently clear. In particular, this holds true for the Kostroma interstadial sediments, stratigraphic position of which needs to be verified.

The upper Neopleistocene of the Russian Plain is represented by deposits of the Mikulinian Interglacial, which are correlative to the Eemian deposits of Central and Western Europe, and by sediments of the succeeding Valdayan cooling phase. Owing to contribution of many researchers, paleobotanical characterization of the Mikulinian deposits is quite adequate (Grichuk, 1982; Velichko *et al.*, 1993). In some palynological diagrams showing two abundance peaks of broad-leaved

[3] Following Velichko (1975) and Velichko *et al.* (1989), other authors of this work interpreted them as two stages of a single glaciation.

taxa at that time, the second one is related to redeposition of corresponding palynomorphs (Grichuk, 1989).

Comprehensive paleogeographic studies of upper Neopleistocene sequences in the central Russian Plain revealed the Kalinin till of the early Valdayan time and the Ostashkovo till of the late Valdayan time. Distribution areas and independence degree of relevant glacial events are subjects of controversy. At least two cryogenic and two warming phases took place in the early Valdayan time. The earliest Kurgolovskii cooling event was followed by the Verkhnyaya Volga warming that is correlative with the Brörup Interstadial, the Amersfoort phase included.

According to some data (Zarrina, 1991; Alekseev *et al.*, 1997; Sudakova *et al.*, 1997), the early Valdayan glaciation left behind tills in the Rzhev, northern Moscow, Rostov, and Rybinsk–Tutaev areas of the Volga River basin. In the Loshakovo, Dmitrov, Cheremoshnik, Chermenino, and Dolgopolka key sections, the Kalininan till of peculiar structure and composition rests immediately on reliably dated Mikulinian sediments (*Putevoditel'...*, 1981; Lazukov *et al.*, 1982; Sudakova *et al.*, 1996; and others). The time span of the Kalininan glaciation is restricted by a series of thermoluminescence dates (Alekseev *et al.*, 1997; Sudakova *et al.*, 1997).

According to alternative viewpoint shared by Pisareva, the early Valdayan glaciation did not advance far beyond the Baltic shield slopes (Velichko *et al.*, 1993). Nevertheless, the relevant cooling phase inferable from all paleobotanical data was significant enough to result in loess accumulation and cryogenic deformations not only in northern, but also in central areas of the East European plain (Velichko *et al.*, 1999).

The middle Valdayan megainterstadial (about 50 ka ago) that comprised the alternating cooling and warming phases is well established based on various data. Climatic conditions at that time were cooler in general than the modern ones.

The progressing cooling resulted in development of glaciation of the late Valdayan time, when glaciers advanced from the Scandinavian glaciation center. Outside the glaciation area there was an extensive permafrost area (Velichko *et al.*, 1999).

The suggested scheme of glacial events development in space and with time (Fig. 2) is elaborated for the entire glaciation area of the Russian Plain (Sudakova, 1993; Sudakova *et al.*, 1998). It leads to the following conclusions.

The glaciation rhythms regular during the Brunhes Chron are traceable at least since the early Eopleistocene. Seven glacial complexes of the Neopleistocene time have been established based on till horizons and other diagnostic features. The inferable distribution limits of peak glaciations suggest that the ice-sheet area tended to reduce from the early Don to the late Valdayan epochs of the Neopleistocene.

In contrast, distribution areas of problematic ice sheets of the pre-Don period probably grew with time, though they did not extend outside the northern and central parts of the Russian Plain. However, these glaciations and their distribution limits remain enigmatic, because subsequent exaration and erosion destroyed their traces on the largest part of the territory.

The extreme progradation to the south is established for ice covers of the Don and Okaian times. The stratigraphic position of the Don Glaciohorizon is well substantiated, whereas the corresponding glaciation center and possible stages of its development are subjects of discussion. Type sections of the Okaian till are widespread in upper and middle courses of the Oka River (the Likhvinka, Bryan'kovo, Tarusa, Narovatovo and other localities).

The marker Dnieper and Moscow glaciohorizons deposited in the mid-Neopleistocene time are distinguished based on all diagnostic criteria. Thorough lithological investigations in large glaciation areas provided grounds to reconstruct the flow and marginal structures of the Dnieper and Moscow glaciers. As mentioned above, lithological characteristics of both tills substantiated their confident correlation over the vast territory from the upper Oka and Protva river basins to the Timan and Cis-Urals (Andreicheva *et al.*, 1997).

In the right-bank areas of Volga River upper courses, there is convincing evidence that glacier of the Kalininan time advanced here in the late Neopleistocene (Sudakova *et al.*, 1998). This does not mean however that two glacial sheets of this period developed in a similar way everywhere around their perimeter.

Paleolandscape reconstructions. Paleolandscapes have been reconstructed for the Moscow and Tver' regions between 56° and 57°N, characteristic of which now are the temperate continental climate and broad-leaved–spruce forests. Many key and stratotype sections comprehensively studied at present are located in this zone. Paleoclimate and biotic components of landscapes are reconstructed based on paleobotanical, faunal, and lithological data. Summarizing paleobotanical data, we paid main attention to those, which characterize most complete sections with lacustrine–boggy deposits most suitable for paleogeographic interpretation. Despite the lack of data on some intervals and very general characteristics of successive vegetation stages (Fig. 2), the principal trend of natural environment development can be rather distinctly outlined.

The Eopleistocene–Early Neopleistocene

During the climatic optimum of the earliest Akulovo Interglacial, the Moscow area was covered by polydominant mixed conifer–broad-leaved forests with diverse fern taxa *Osmunda cinnamomea, O. claytoniana, O. regalis, Onoclea* sp., and *Adiantum* sp. Composition of the fossil flora suggests moderately warm cli-

matic conditions with average temperature 0°C in January and about +22°C in July.

The independent status of the next Okatovo Interglacial is proved by the preceding climatic pessimum. The temperature drop during the latter and continental character of climate are inferable from increasing abundance of herbaceous and fruticose pollen taxa and from decreased proportion of arboreal pollen lacking representatives of thermophilic species. Palynological spectra are dominated by boreal taxa associated commonly with specimens of tundra and arctic–alpine vegetation. Such a composition of plant community is characteristic of the forest–tundra landscapes.

Five stages of vegetation development have been established for the Okatovo Interglacial. During the climatic optimum, polydominant broad-leaved forests in western and northern areas of the Moscow region consisted of oak, elm, lime, and hornbeam with admixture of wing-nut, chestnut, *Cornus*, grape, and Myricaceae (Fursikova *et al.*, 1992; Shik *et al.*, 1994).

The periglacial forest–steppe appeared at the degradation time of the Don ice cover (Pisareva *et al.*, 1979). Frequent finds of *Alnaster fruticosus* are indications of permafrost conditions in a vast territory, the Don River basin included.

There were three climatic optima during the long-lasted Roslavl' (Muchkap) Interglacial. The earliest Glazovo optimum was characterized by prevalence of polydominant broad-leaved forests, which advanced northward to nearly 59° N and southward to 51° N. Broad-leaved–coniferous forests with subordinate hornbeam, which appeared during the second Konakhovka optimum, were replaced afterward, in the third Galich optimum, by coniferous–broad-leaved forests with fir and hornbeam. Boreal vegetation dominated occasionally by spruce and spruce–pine forests locally swamped was typical of the intervals between the optima (Pisareva, 1997). During degradation of the Okaian glaciation, landscapes represented a periglacial forest–tundra.

Middle Neopleistocene

In the Likhvinian Interglacial, vegetation in upper courses of the Volga River was dominated by pine forests with fir and subordinate oak, hornbeam, wing-nut, chestnut, and grape. Continuous sections in the Tver' and northern Moscow areas recorded deterioration of climate after the Likhvinian Interglacial that is evident from development of boreal forest formations at that time. Traces of periglacial vegetation during the first post-Likhvinian cooling are unknown.

Natural environments changed at the time of the Kosha Interstadial, as it was established first by Grichuk (1989) in the Bol'shaya Kosha section and recognized later on in the lacustrine sequences of the upper Volga and Moscow areas. According to paleobotanical data, the Kosha vegetation resembled the present-day taiga of West Siberia. Composition of fossil flora suggests that the temperature was not higher than –16°C in the coldest month and +17°C in the warmest one.

The subsequent landscape evolution was influenced by progressive climatic cooling and transition to more continental climate that resulted in development of light birch forests with larch and periglacial landscape. The Mar'ino Interglacial (established in the section near the eponymous village in the northern Moscow area) was characterized by expansion of conifer and birch forests.

The Kaluga cooling established in the Likhin stratotype section was most significant one in the post-Likhvinian time (*Razrezy...*, 1977; Sudakova, 1975; Bolikhovskaya, 1995). Periglacial tundra of the Kaluga time dominated in the region under consideration. This significant cooling could be induced by development of glaciation in northern areas of the Russian Plain, southern boundaries of which are now under examination. Climate of the following Chekalin Interglacial (*Razrezy...*, 1977) was cooler than that of the Likhvinian Interglacial and stimulated development of mixed conifer–broad-leaved forests of a less diverse composition (Bolikhovskaya, 1995). It is likely that between the Chekalin Interglacial and the onset of the Dnieper glaciation there were frequent climatic oscillations.

Lemming fauna from the Chekalin and Kopaevo (Rybinsk area) sections (Agadzhanyan, 1992) is indicative of periglacial environments before the Dnieper glaciation. Specific biotopes of this period imply that the region in general resembled the tundra landscapes of Yakutia. Many researchers described various structural features, such as ice wedges, potholes, and cryoturbation signs, characteristic of that period and produced by freezing, permafrost conditions, and other cryogenic phenomena.

Available information on the problematic warming between the Dnieperian and Moskowian glaciations is insufficient for a confident evaluation of landscape and climatic changes in the second half of the middle Neopleistocene. Despite different interpretations of the glacial events, paleobotanical data suggest the interstadial conditions between these glaciations. Sparse spruce forests with admixture of fir, cedar, larch, pine, and birch grew during this climatic optimum in central areas of the plain (Pisareva, 1965). This supports viewpoints of Markov (1939, 1955), Velichko *et al.* (1989) and other specialists who argued for the stadial development of the Dnieperian and Moskowian glaciations.

Episodes of warming and less contrast climatic conditions occurred during the Moscow Glacial, climate of which was severe in general. The warming and cooling phases might be related to stages of the ice sheet retreating and advancing. Sparse forests with *Picea obovata* and pine grew in the period separating the Moscow Glacial and the Mikulinian Interglacial. Arctic–boreal and arctic–alpine species present in corre-

sponding flora and typical of periglacial vegetation represent a distinctive feature of that period.

Late Neopleistocene

After works of V.S. Dokturovskii, V.I. Sukachev, V.P. Grichuk, E.M. Zelikson and other researchers, vegetation and climates of the Mikulinian (Eemian) Interglacial have been comprehensively understood. Summarizing the available paleobotanical data, we reconstructed the landscape evolution throughout this interglacial period (Fig. 2). Landscape zones of the Mikulinian time were similar in general to the recent ones, northern and southern limits of broad-leaved forests were displaced northward and southward, respectively, during the warming phases (Velichko *et al.*, 1999).

Abundant biostratigraphic data have been used to distinguish successive environmental phases of the Valdayan Glacial in the European Russia (Borisova and Faustova, 1994; Velichko *et al.*, 1999). Of two cooling phases of the early Valdayan time, the second one was colder, characterized by periglacial vegetation in central areas of the region. With all its warming and cooling events taken into consideration, climate of the middle Valdayan (Monchalovo) period was more severe than the recent one.

In the late Valdayan time, the peak cooling occurred 23–17 ka ago. A severe continental climate of the late Valdayan time caused the complete degradation of forest vegetation. Approximately 18 ka ago, forestless landscapes hosted arctic–alpine, boreal, and steppe species, with halophytes dominated in the Yaroslavl, Ivanovo, and Kostroma areas of the Volga River basin. These climatic conditions enlarged distribution areas of permafrost and cryogenic lithogenesis. Radiocarbon dates for the terminal stage of the late Valdayan deglaciation range from 11 to 10.5 ka.

Paleoceanological records. As is mentioned above, the established secular variations of oxygen isotope parameters in oceanic sequences have been widely used for recalibration of climatostratigraphic schemes with intent to unify both data sets. At first sight, these variations outline in general a regular saw-tooth oscillations of the Neopleistocene climate, and ranking of continental climatic events retreated to the back line.

Regretfully, age determination of Quaternary continental deposits, especially older than ~0.35 Ma, is far from being precise.

Without reliable markers, the subdivision and correlation of continental sections is additionally complicated by different interpretations of regional reconstructions, stratigraphic and paleogeographic schemes, and by insufficiently correct ranking of individual climatic events. Correlation between continental and oceanic paleogeographic events is even much more difficult.

In our team as well there diverging viewpoints on some continental events. Their correlation with oceanic events is aggravated by the fact that the paleomagnetic Brunhes–Matuyama boundary has not been uniformly defined in sections of continental glacial sediments (Faustov *et al.*, 1997).

In fact, the common practice of correlating oceanic and continental paleogeographic events is to count warming and cooling epochs from the top downward. However, we actually do not know the accurately defined and commonly accepted rank and stratigraphic position of a given event and relevant deposits, and this practice leads to subjective results. The way out of the situation seems to be in ranking and grouping of major climatic events based on oceanic and continental data, or in using of some appropriate markers in addition to the Brunhes–Matuyama boundary, as it has been done in some works (Kukla, 1977; Kukla and Cilek, 1996; Zubakov, 1986). To correlate oceanic and continental sequences without additional markers is very difficult.

Mathematical processing of oxygen-isotope variation curves showed the meridional nonequivalence of $\delta^{18}O$ amplitude variations in carbonates of the same isotopic stages, though four principal peaks are permanently distinguishable everywhere. In addition, it was established that intensity of the main climatic events varied with time in different latitudinal—climatic zones of the World Ocean (Nikolaev, 2000). Of prime significance are isotopic stages 2 and 6 reflecting the cold climatic events of the terminal Brunhes Chron, which are correlative with the late Würm and Riss (late Riss?) glaciations of the Alpine scale. These were most intense climatic events in all latitudinal zones except for the polar one, where the cooling stage 6 is poorly distinguishable. Most significant cooling events of the earlier period correspond to different isotope stages depending on latitudinal–climatic zones.

In the equatorial zone, peaks of oxygen isotope stages 16 and 22 are as intense as those of stages 2 and 6, whereas that of stage 14 was minimal. In the tropical zone, most significant peaks indicative of cooling events during the first half of glaciation epoch in the Northern Hemisphere correspond to stages 16 and 20 (instead of stage 22).

In the next northerly zone, the main cooling peak is that of stage 16 in the first half of the Brunhes Chron. Farther to the north, at the latitude of the southern extreme of Greenland, maximum peaks mark cold stages 12 and 16. At 63° N, the greatest peaks characterize stages 12 and 14. In the Arctic zone (80° N), the coldest climatic events are recorded at stages 8 and 12, and previous cooling peaks are less intense. The observable variability reflects most likely the variable contribution of temperature-dependent components in the isotopic signal, because the stirring period of oceanic waters does not exceed 1.5 kyr. The established latitudinal dependence of cooling peak intensities recorded in oceanic sediments can be regarded as suit-

able explanation for metachronism of glaciations in terminology of Markov. It may be helpful for correlation of climatic events.

For the comparison with continental data, we used two curves of oxygen isotope variations in planktonic foraminifer tests (Fig. 2), which have been established at the ODP Sites 919 (63° N and 37° W) and 646 (56° N and 48° W). Both sites are situated in the latitudinal zone of Atlantic Ocean that is comparable with the continental zone under consideration.

In distinction from the first curve, the second one characterizing the ODP site located closer to the south is better correlative with dynamics of continental landscape evolution shown in the same figure. Two main peaks of the first curve are those of oxygen isotope stages 12 and 14, which we correlate with the significant Okaian and Don glaciations. As the distribution area of Don till extends far to the south, the curve established at the lower latitude shows an additional peak corresponding to stage 8.

In the curve depicting dynamics of continental landscape zones, the first half of the Brunhes Chron is marked by two coldest climatic episodes of the Okaian and Don times, which are separated by the warmest Muchkap epoch. Simultaneously the southern isotopic curve shows, in distinction from the northerly one, a significant cooling peak of stage 8 (it is correlative with the Dnieper Glacial), the lesser one of stage 14, and the noticeable warming event of stage 11 (the Likhvinian Interglacial).

If the suggested correlation of first-order rhythms is correct, we may try to correlate the shorter-term rhythms. Using data of Bassinot *et al.* (1994) as a basis for comparison, we will start the correlation procedure from the top downward, because younger deposits are naturally studied in more detail. However, as the chosen isotopic curves characterize oceanic sequences of lower latitudes, one could hardly expect the analogous intensity of correlative peaks.

We found a rather good covariation of both curves under comparison down to the stage 3. The coincidence is incomplete however, because the minor Kashino warming is not distinguishable in the isotopic variations, the known SPECMAP curve included. Because the correlation between the middle Valdayan warming and stage 3 is beyond doubt, we can conclude that the continental record is of a greater sensitivity.

The complete coincidence of curves is observable within the interval of stage 5. There are three cooling peaks in stage 6, and four significant cooling episodes of the Moscow time are recorded in continental sequences. In distinction from three warming peaks characterizing stage 7, continental data depict a single warming event, though continental sections of corresponding age are incomplete. The Dnieper till of three-layer structure is comparable with short-term oscillations within stage 8 of the SPECMAP curve. Among two peaks of stage 9, one is insignificant, and the whole stage can be correlated with the indivisible Chekalin Interglacial. Two cooling phases of the Kaluga Glacial correspond to the peak couple of stage 9. Minor rhythms of Stage 10 coincide with those of the Likhvinian Interglacial. Bassinot and his colleagues depicted a single peak of stage 12 that is correlative with the Okaian Glacial. The triple peak of Stage 13 corresponds to three warm phases of the Muchkap Interglacial. Further correlation of second-order rhythms is useless, because necessary continental records are fragmentary. Nevertheless, as is shown above, the correlation between short-term climatic changes in continental and oceanic areas is plausible within the interval down to oxygen isotope stage 13 inclusive.

CONCLUSIONS

1. The revision of Neopleistocene sequences in glaciation areas of the Russian Plain showed that the Brunhes Chron can be confidently divided into eight large climatostratigraphic rhythms (Setun–Okatovo, Don–Roslavl', Oka–Likhvinian, Kaluga–Chekalin, Dnieper–subsequent interstadial, Moscow–Mikulinian, early Valdayan–middle Valdayan, and late Valdayan–Holocene pairs of cooling and warming events). At least seven glaciations distinguishable within this time span left behind till horizons and other characteristic diagnostic features.

2. Paleobotanical data enable recognition of successive stages in vegetation evolution during the early Neopleistocene–Holocene period. They show that interglacial and some glacial epochs included complex series of paleogeographic events.

3. Comparing intensities of climatic signals recorded in corresponding continental and oceanic latitudinal zones, we found possibility to correlate the Valdayan Glacial with oxygen isotope stages 2 and 4, the Moscow Glacial with stage 6, the Dnieperian glaciation with stage 8, those of the Okaian and Don times with stages 12 and 14, respectively, and the Setun' Glacial with stage 16.

4. The correlation between high-frequency rhythms turned out to be plausible down to the oxygen isotope stage 13 inclusive. As is shown, the isotopic substages are well correlative climatic oscillations within the first-order rhythms distinguishable on continent.

ACKNOWLEDGMENTS

This work was supported by the Russian Foundation for Basic Research, project no. 02-05-64879.

Reviewers A.A. Velichko and Yu.A. Lavrushin

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